THE HOLOCENE PALEOLIMNOLOGY OF LAKE SALPETÉN, GUATEMALA

By

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Just because some of us can read and write and do a little math, that doesn't mean we deserve to conquer the universe.

Kurt Vonnegut, Jr.

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TABLE OF CONTENTS

Abstract of Dissertation Presented to the Graduate School
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Stratigraphic shifts in the oxygen isotopic ($\delta^{18}O$) composition of biogenic carbonate from tropical lake sediment cores are often interpreted as a proxy record of the changing relation between evaporation and precipitation (E/P). Holocene $\delta^{18}O$ records from Lake Salpetén, Guatemala, were apparently affected by changes in drainage basin vegetation that influenced watershed hydrology, thereby confounding paleoclimatic interpretations. Oxygen isotope values in the lake were greatest between ~9900 and 7500 cal yrs B.P., suggesting relatively high E/P, but pollen data indicate moist conditions and extensive forest cover in the early Holocene. The discrepancy between pollen and isotope-inferred climate conditions may be reconciled if the high early Holocene $\delta^{18}O$ values were controlled principally by low surface run-off and groundwater flow to the lake, rather than high E/P. Dense forest cover in the early Holocene would have increased evapotranspiration and soil moisture storage, thereby reducing delivery of meteoric water to the lake.

Biogenic carbonate $\delta^{18}O$ decreased between 7500 and 3300 cal yrs B.P. in Lake Salpetén. This decline coincided with palynologically documented forest loss that may have led to increased surface and groundwater flow to the lake. Minimum $\delta^{18}O$ values occurred between 2400 and 1800 cal yrs B.P. Relatively high lake levels were confirmed by radiocarbon dated aquatic gastropods from subaerial soil profiles ~1.0 to 7.5 m above present lake stage. High lake levels were a consequence of lower E/P and/or greater surface runoff and groundwater inflow related to watershed deforestation by the Maya. When the Maya population declined ~1200 cal yrs B.P., $\delta^{18}O$ values increased as a consequence of reduced hydrologic input caused by increased E/P or forest recovery.

Model simulations incorporating pollen-derived deforestation rates can account for nearly 75% of the variance in the Lake Salpetén biogenic carbonate $\delta^{18}O$ record, but do not reproduce the full magnitude of the $\delta^{18}O$ minimum nor the abrupt changes observed in sediment cores. Minimum $\delta^{18}O$ values were achieved only through combination of both a protracted ~15% increase in precipitation and pollen-based estimates of vegetation cover change. Simulation of abrupt $\delta^{18}O$ changes necessitated the rapid onset of precipitation decreases of at least 10% relative to modern.

CHAPTER 1
INTRODUCTION

Instrumental records of climate span only the last two centuries. Paleo-environmental methods are therefore required to assess pre-instrumental conditions in terrestrial and aquatic systems and to evaluate whether natural or anthropogenic factors exert primary control on the structure and function of ecosystems. Lake sediment cores have been used extensively to decipher the history of aquatic ecosystems and surrounding watersheds. Sediment profiles record both long-term, climate-driven environmental changes and the results of recent anthropogenic impacts. These archives thus provide insights into the magnitude of human-mediated environmental shifts.

This study investigates Holocene environmental change in the Maya lowlands of Petén, Guatemala. The occupation of the Maya Lowlands over the last 3000 years serves as a large-scale experiment in the human use of a tropical karst environment. The results of this pre-historic experiment are recorded in lake sediments and can be deciphered through multi-disciplinary study of lacustrine deposits. I used stratigraphic variations in the oxygen isotopic composition ($\delta^{18}O$) and trace element composition of biogenic carbonate from Lake Salpetén sediment cores to infer past environmental changes in the region. These sediment variables indicated that pronounced changes in watershed hydrologic balance were caused by both human-induced deforestation and natural climate change. Quantitative modeling of the basin provides further insight into the hydrologic and isotopic response of the lake to natural and anthropogenic disturbances.

Isotope-Based Paleoenvironmental Studies

There are three naturally occurring stable isotopes of oxygen, ^{16}O, ^{17}O, and ^{18}O, with relative natural abundances of 99.7630%, 0.0375%, and 0.1995%, respectively. Because of their mass differences, the isotopes exhibit subtle differences when they enter into physical, chemical, and biological processes in the environment. It is the differential behavior of the ^{18}O and ^{16}O isotopes and their changing relative abundance in the environment that are exploited for paleoclimate reconstructions. The oxygen isotopic composition ($\delta^{18}O$) of environmental samples can be measured by isotope ratio mass spectrometry and reported as the ratio of ^{18}O to ^{16}O in standard delta (δ) notation.

$$\delta^{18}\mathrm{O} = \left[\frac{\left(^{18}\mathrm{O} / {}^{16}\mathrm{O}\right)_{\mathrm{sample}} - \left(^{18}\mathrm{O} / {}^{16}\mathrm{O}\right)_{\mathrm{standard}}}{\left(^{18}\mathrm{O} / {}^{16}\mathrm{O}\right)_{\mathrm{standard}}} \right] \times 1000$$

Standards used for oxygen in lacustrine carbonates and water samples are the Pee Dee Belemnite (PDB) and Standard Mean Ocean Water (SMOW) respectively (Craig 1961).

Several studies have reviewed the rationale for using the stable isotope ($\delta^{18}O$) signal in carbonate shells from lake sediment cores to reconstruct past climate conditions (Talbot 1990; Curtis and Hodell 1993; Holmes 1996; Schwalb 2003). Variations in the oxygen isotopic ratio ($^{18}O/^{16}O$) in sedimented shell material result from variations in the temperature at which carbonate precipitation occurred, the isotopic composition of the lake water at the time the organism lived, and biological fractionation by the organism (von Grafenstein et al. 1999). Because there is little evidence for major excursions in the mean temperature of the tropics during the Holocene, the changing $\delta^{18}O$ of lake water over the last 10,000 years has been the major determinant of the $\delta^{18}O$ in shell carbonate.

In lakes with negligible surface outflow, variations in the $^{18}O/^{16}O$ of lake water are most influenced by changes in the rate of evaporation (E) relative to the combined inputs

of precipitation (P) and surface and groundwater inflow (I) (Fontes and Gonfiantini 1967). During extended dry periods (high E/P or E/I) ^{18}O becomes relatively concentrated in the lake water because $H_2{}^{16}O$, with its higher vapor pressure, is preferentially lost to evaporation. Conversely, increased surface and groundwater inflow or precipitation results in lower $^{18}O/^{16}O$ ratios.

Short-lived aquatic organisms which form calcium carbonate shells preserve a record of the E/P ratio that prevailed during their lifetimes. When they die, their remains are buried in sediments on the lake bottom, thereby preserving an archive of past climate change. The stratigraphic paleoclimate record can be deciphered by mass spectrometric measurement of the $\delta^{18}O$ of sedimented shells. More positive $\delta^{18}O$ values generally indicate higher E/P (drier) conditions and/or decreased surface and groundwater inflow, whereas more negative values indicate relatively lower E/P (moister) conditions and/or increased inflow at the time the organism lived.

Environmental Setting

The Yucatán Peninsula includes the Department of Petén in northern Guatemala, Belize, and the Mexican states of Campeche, Yucatán, Quintana Roo, and portions of eastern Tabasco and Chiapas (Wilson 1980) (Fig. 1a). The region is characterized principally by low-lying karst topography developed on Cretaceous and Tertiary limestones. In the northern lowlands, karstic terrain is fully developed and surface drainage is virtually non-existent. Surface elevation is relatively low, however, and the water table is exposed in the many sinkholes (*cenotes*) where the limestone has collapsed. In the southern lowlands of Petén, surface elevation varies between 100 and 500 m above mean sea level and groundwater is not readily accessible. Karst landforms are less developed and surface waters are perched, resulting in numerous stable lakes and

seasonally inundated topographic depressions (Deevey et al. 1979). The central lake district contains a series of terminal basins centered at 17°N and distributed along a system of east-west aligned faults (Fig. 1b).

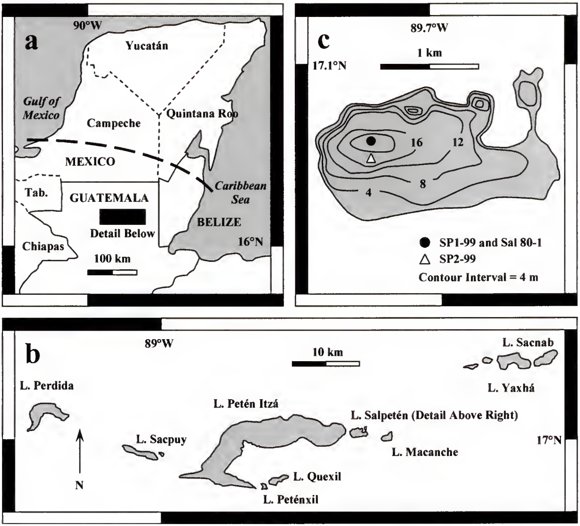

Figure 1-1. Lake study sites in the Department of Petén, Guatemala. (a) Map of the Yucatán Peninsula showing the location of the Petén lake district in northern Guatemala (black rectangle) and Lakes Chichancanab (1) and Punta Laguna (2). Heavy dashed line indicates the division between the northern and southern Maya lowlands. (b) Petén lake district in Guatemala. (c) Bathymetry and core sites in Lake Salpetén.

Annual rainfall across the Yucatán Peninsula is highly variable, ranging from a low of 500 mm yr^{-1} along the northwest coasts of Yucatán and Campeche to over 4000 mm yr^{-1} in the southern lowlands (Wilson 1980). In Petén, rainfall varies locally and annually from ~900 to 4000 mm, with a regional mean of 1600 mm (Lundell 1937). The distribution of this rainfall throughout the year is highly seasonal. Dry conditions develop in November and December as the inter-tropical convergence zone (ITCZ) and Azores-Bermuda high-pressure system move toward the equator and strong trade winds become predominant (Portig 1976; Hastenrath 1976; 1984). Intense convection associated with the northward migration of the ITCZ and the Azores-Bermuda High gives rise to the heavy rains over the region between June and October when the trade winds tend to be weak and warm sea-surface temperatures prevail in the tropical Atlantic.

The effects of seasonal aridity and the precipitation gradient are reflected in the soil development and vegetation of Yucatán. Soils in the northern lowlands are thin and support dry-adapted scrub vegetation and some semi-evergreen forest of medium height (Wilson 1980). Farther south, in Petén, the landscape is dominated by well-drained forest soils and tropical semi-deciduous and evergreen vegetation (Lundell 1937). The forest soils of Petén are highly fertile, and only high susceptibility to erosion and presence on steep slopes restrict cultivation (Deevey et al. 1979; Rice et al. 1985). Under conditions of deforestation, enhanced transport of dissolved and particulate matter is expected, and the tremendous erosive potential of intense tropical rain is a major contributing factor in the transfer of materials from the land to the lake bottom.

Maya Prehistory of the Yucatán Peninsula

Maya populations have occupied the lowlands of the Yucatán Peninsula from Early Preclassic times, beginning ~4000 years ago (Hammond 1976; 1979). By 250 cal

yrs A.D., the Maya lowlands were densely populated with well-developed urban nuclei. Late Classic (550 to 830 cal yrs A.D.) population estimates for the southern lowlands, for example, vary between three and fourteen million people (Thompson 1966; Adams et al. 1981; Rice and Puleston 1981; Turner 1990). The urban site of Tikal alone may have supported a Late Classic population in excess of sixty-thousand (Haviland 1969; 1972; Culbert et al. 1990). But in the 9th century AD, Classic Maya civilization suffered a near-total collapse, evident in a complete cessation of the arts and architecture associated with the ruling class and in a dramatic reduction of much of the regional population (Lowe 1985; Messenger 1990; Rice and Culbert 1990; Sharer 1994). To the north, population decline was less dramatic, and the Postclassic Period (900 to 1525 cal yrs A.D.) witnessed a significant resurgence (Dahlin 1983; Folan et al. 1983; Rice and Culbert 1990; Gill 2000). Sites in the southern lowlands, however, show minimal evidence of reoccupation (Gunn and Adams 1981; Folan et al. 1983; Rice and Culbert 1990).

Maya colonization of the Petén dates from the Middle Preclassic Period (800 to 300 cal yrs B.C.) (Rice and Rice 1990). By the Late Preclassic (300 cal yrs B.C. to 250 cal yrs A.D.), as Maya settlement and cultivation increased, much of the Petén was deforested (Cowgill and Hutchinson 1966; Deevey et al. 1979). However, Late Preclassic and Early Classic population growth was low in the central Petén lake district (Rice and Rice 1990). Only the eastern Petén basins of Yaxhá and Sacnab experienced accelerated growth in the Early Classic (Culbert et al. 1990). In contrast, population growth during the Late Classic was dramatic in all basins of the central Petén, with the highest settlement densities (250 individuals km^{-2}) occurring in the catchments of Lakes Salpetén and Macanche (Rice and Rice 1990). The Terminal Classic period, consistent

with the decline of the lowland Maya civilization, reflects uniform decline in settlement throughout all of the central Petén lake basins. The eastern Petén lake basins experienced a minor resurgence of population during the Early Postclassic, but population declined consistently throughout the Late Postclassic. Maya populations were limited primarily to the area of Lake Petén Itzá at the time of Spanish contact (Schwartz 1990).

Previous Paleoenvironmental Studies In Petén

Previous paleoenvironmental studies in lowland Petén, Guatemala documented Pleistocene/Holocene vegetation changes and long-term human impacts on terrestrial and aquatic systems (Cowgill and Hutchinson 1966; Deevey et al. 1979; Leyden 1984; Rice et al. 1985; Vaughan et al. 1985; Binford et al. 1987; Leyden et al. 1993; 1994). Lake Petén Itzá provides a comprehensive multi-proxy record of Holocene paleoenvironmental change in the central lake district of northern Guatemala (Islebe et al. 1996; Curtis et al. 1998). Oxygen isotopic records of biogenic carbonate from Lake Petén Itzá indicate lake basin filling by ~9000 radiocarbon years before present (^{14}C yrs B.P.), coincident with the filling of other lakes in Petén and shallow basins farther north on the Yucatán Peninsula. Relative aridity during the early Holocene is inferred from the $\delta^{18}O$ record, but contrasts with regional pollen evidence of widespread moist tropical forest by 8500 ^{14}C yrs B.P. (Leyden 1984; Vaughan et al. 1985; Leyden 1987; Leyden et al. 1993; Islebe et al. 1996). The $\delta^{18}O$ values from Petén Itzá display a marked decrease between about 7000 and 6000 ^{14}C yrs B.P., reflecting a transition to moist conditions. The early Holocene $\delta^{18}O$ record from Lake Petén Itzá is difficult to interpret, however, because the basin was in the early stages of filling. Discrepancy between early Holocene pollen and $\delta^{18}O$ results from Petén Itzá can be reconciled if lake water $\delta^{18}O$, strictly interpreted as

changing E/P, was controlled not only by regional climate but by the changing surface area to volume ratio of the filling lake (Curtis et al. 1998).

After 5000 ^{14}C yrs B.P., the $\delta^{18}O$ record from the basin demonstrates little variability (Curtis et al. 1998). Rather than indicating climate stability in the late Holocene, the $\delta^{18}O$ record from Lake Petén Itzá may simply reflect the insensitivity of large volume lakes to short-term changes in E/P. Unfortunately, a late Holocene climatic inference based on other proxies is confounded by human impact on catchment vegetation and soils (Curtis et al. 1998). For example, the pollen record from Lake Petén Itzá provides evidence for the decline of tropical forest and increased disturbance taxa between ~4000 and 1000 ^{14}C yrs B.P. (Islebe et al. 1996). The decline of forest taxa in the Petén Itzá record is consistent with regional forest clearance by Maya populations (Cowgill and Hutchinson 1966; Deevey et al. 1979; Vaughan et al. 1985; Leyden 1987). Sediment magnetic susceptibility and elemental geochemistry suggest that forest clearance accelerated the erosion of catchment soils. Erosional inputs to the lake began to subside by ~1100 ^{14}C yrs B.P., following the collapse of Classic Maya civilization (Curtis et al. 1998). Forest regeneration, evident in increased preservation of forest pollen and decreased disturbance taxa, dates from ~1000 ^{14}C yrs B.P. (Islebe et al. 1996).

Statement of Objectives

In July 1997 and August 1999, sediment cores were recovered from six additional lakes in the Department of Petén, northern Guatemala. Initial efforts were focused on sediment profiles from Lake Salpetén, a small water body of relatively high salinity (4500 mg l^{-1} TDS) (Figure 1-1c). Salpetén is an exceptional site for paleoenvironmental research. Evaporative loss from the lake is substantial and temporal changes in the

$^{18}O/^{16}O$ ratio of the lake water are controlled primarily by changes in the ratio of evaporation to input (E/I). Significant evaporative loss is demonstrated by the fact that lake waters are enriched relative to precipitation by 4 to 5 ‰. Past changes in the $^{18}O/^{16}O$ ratio of the lake water are therefore recorded in the continuous $\delta^{18}O$ record of carbonate in gastropod and ostracode shells preserved in the sediments. The following chapters explore the interpretation of the oxygen isotopic date from biogenic carbonates from Lake Salpetén. Chapters 2 through 4 are intended for publication and therefore contain redundancies in the introductory text.

CHAPTER 2
INFLUENCE OF VEGETATION CHANGE ON WATERSHED HYDROLOGY: IMPLICATIONS FOR PALEOCLIMATIC INTERPRETATION OF LACUSTRINE $\delta^{18}O$ RECORDS

Introduction

Watershed hydrology and the related transfer of materials from terrestrial to aquatic systems are a function of several factors, including the lake/watershed ratio, lake morphometry, local climate variables and watershed characteristics such as bedrock geology, soil type, topography, and vegetation cover (Mason et al. 1994; Street-Perrott 1995). Removal of vegetation as a result of climate change or clearance by humans can reduce transpiration and soil moisture storage, thereby increasing water and material transfer to a lake or groundwater (Hibbert 1967; Bosch and Hewlett 1982; Bruijnzeel 1990; Hornbeck and Swank 1992; Stednick 1996). Rates of water and material transfer to lacustrine systems, in turn, strongly influence in-lake biogeochemical processes. Historical changes in the delivery rates of dissolved and particulate materials can often be detected in accumulated lake sediments. Reliable paleoenvironmental reconstructions based on lacustrine sediment records therefore require an understanding of all factors that affect catchment processes, including human disturbance (Frey 1969; Oldfield 1978; Pennington 1981; Binford et al. 1983; Deevey 1984; Binford et al. 1987).

Previous paleoenvironmental studies in lowland Petén, Guatemala (Figure 2-1a), documented Pleistocene/Holocene climate and vegetation changes and long-term human impacts on terrestrial and aquatic systems (Cowgill and Hutchinson 1966;

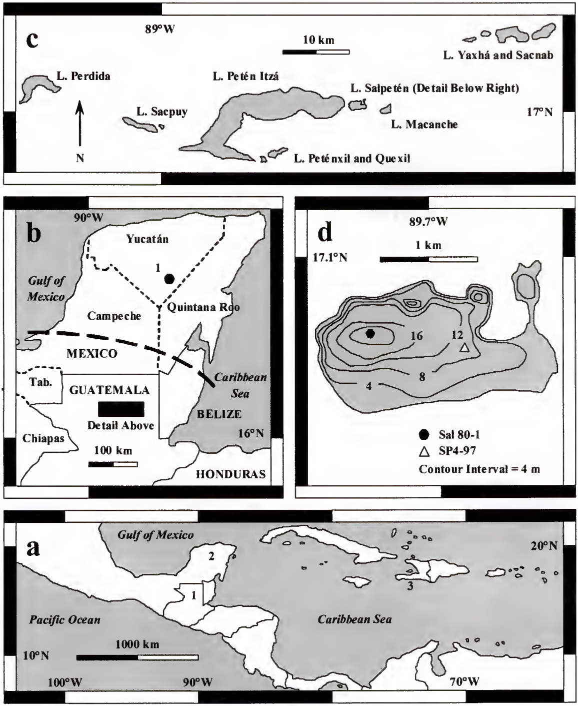

Figure 2-1. Lake study sites in the Caribbean region. (a) Map of the Caribbean showing lake study sites in northern Guatemala (1) and Lakes Chichancanab (2) and Miragoane (3). (b) Map of the Yucatán Peninsula showing lake study sites in northern Guatemala (black rectangle) and Lake Chichancanab (1). Heavy dashed line indicates the division between the northern and southern Maya lowlands. (c) Petén lake district in Guatemala. (d) Bathymetry and sediment core sites in Lake Salpetén.

Dahlin 1983; Leyden 1984; Rice et al. 1985; Vaughan et al. 1985; Binford et al. 1987; Leyden et al. 1993; Islebe et al. 1996; Curtis et al. 1998). In this study, Holocene environmental change in the Lake Salpetén watershed was inferred from multiple sediment cores. Stratigraphic variations in the oxygen isotopic ($\delta^{18}O$) and trace element (Mg and Sr) composition of biogenic carbonate from these cores were interpreted as reflecting altered watershed hydrologic balance that resulted from changes in climate, vegetation, and human disturbance in the catchment. Changes in material transfer to the lake bottom were inferred from sediment composition and accumulation, and vegetation changes within the catchment were reconstructed from fossil pollen assemblages. The Lake Salpetén record was compared with a profile from nearby Lake Petén Itzá (Curtis et al. 1998) and sediment records from Lake Chichancanab, Mexico (Hodell et al. 1995) and Lake Miragoane, Haiti (Hodell et al. 1991; Curtis and Hodell 1993).

Proxy Indicators of Environmental Change

Variations in the oxygen isotopic ratio ($^{18}O/^{16}O$) of lacustrine carbonate result from changes in the temperature of carbonate precipitation and shifts in the $^{18}O/^{16}O$ composition of the lake water from which the carbonate precipitates (Craig 1965; Gonfiantini 1965; Stuiver 1968; 1970; Covich and Stuiver 1974; Fritz et al. 1975). In tropical lakes, temperature fluctuations are assumed to be minor during the Holocene and variations in the $^{18}O/^{16}O$ of lacustrine carbonate are thought to have been dominated by changes in lake water residence time and the relative rates and isotopic composition of hydrologic inputs and outputs (Gasse et al. 1990; Talbot 1990; Hodell et al. 1991; Lister et al. 1991; Curtis and Hodell 1993; Curtis et al. 1996; Xia et al. 1997).

The hydrologic balance of a lake (dV_{LAKE}) is controlled by the transfer of water to and from the catchment according to the equation:

$$dV_{LAKE} = \Sigma I + P - \Sigma O - E$$

where ΣI and ΣO are the total surface and groundwater inflows (I) to, and outflows (O) from the lake, P is direct precipitation on the lake, and E is the evaporative loss from the lake (Pearson & Coplen, 1978; Gat, 1984). A similar expression can be written for the oxygen isotopic composition of the lake water:

$$dV_{LAKE}\,\delta_{LAKE} = \Sigma I\,\delta_I + P\,\delta_P - \Sigma O\,\delta_O - E\,\delta_E$$

where δ is the isotopic composition of the various inputs and outputs. In lakes with negligible surface outflow (O), the water balance and lake water isotopic composition is determined by the difference between evaporation and precipitation over the lake (E–P) and the water balance of the surrounding catchment (ΣI) (Craig 1965; Gonfiantini 1965; Fontes and Gonfiantini 1967; Phillips et al. 1986). Lake water ^{18}O enrichment relative to the isotopic signature of direct precipitation and surface and groundwater inflow reflects preferential evaporative loss of $H_2{}^{16}O$ from the lake. Extended periods of enhanced evaporation or reduced precipitation and surface inflow (high E/P or E/I) result in higher $^{18}O/^{16}O$ ratios of lake water and precipitated carbonate. Conversely, increased surface inflow or precipitation (low E/P or E/I) results in lower $^{18}O/^{16}O$ ratios.

Trace element (especially Mg and Sr) chemistry of ostracode valves also serves as an indicator of past changes in watershed hydrology (Chivas et al. 1993; Holmes 1996). Decreased rainfall or surface flow results in enhanced precipitation of lacustrine carbonate and trace element enrichment of lake water. The magnesium and strontium concentration of biogenic carbonate in turn reflects the composition of the lake water and

thereby the hydrologic balance of a lake (Chivas et al. 1985; Chivas et al. 1986). Incorporation of magnesium in carbonate is also affected by water temperature (increased temperature results in increased Mg concentrations) but concentration in lake water is likely to be the major control in tropical lakes.

In addition to altering the hydrologic budget of a lake, increased or decreased precipitation and surface flow alter the transfer of dissolved and particulate materials to a lake. The flux of materials to the sediments is a function of the rate of material output from the catchment (Imboden and Lerman 1978; Binford et al. 1983; 1987). Changes in the material input to a lake produce stratigraphic changes in both sediment accumulation rate and the biogeochemistry of sediment (Deevey 1984; Engstrom and Wright 1984).

Changes in lacustrine sediment composition arise primarily from variations in material loss from the catchment, often the result of land-use changes. For example, soil changes resulting from deforestation and fires produce variations in the flux of particulate matter to a lake (Dearing et al. 1987; Walling 1988; Dearing 1991). Forest clearance and high rainfall accelerate alluviation and colluviation, and increases in sediment accumulation may reflect intensified surface flow and erosion within a watershed. Catchment deforestation, agricultural production, field abandonment, and subsequent plant succession are reflected by the pollen deposited in lake sediments. Deforestation enhances transport of soil nutrients and organic and inorganic matter to the lake, which is reflected in the sediment lithologic composition (Deevey 1984; Binford et al. 1987).

Environmental Setting

The Yucatán Peninsula includes the Department of Petén in northern Guatemala, Belize, and the Mexican states of Campeche, Yucatán, Quintana Roo, and portions of Tabasco and Chiapas (Wilson 1980) (Figure 2-1b). In the northern lowlands, karstic

terrain is fully developed, surface drainage is virtually non-existent, and water falling on the land surface is evaporated, transpired, or quickly delivered to the aquifer. Surface elevation is relatively low, however, and the water table is exposed in the many sinkholes where the limestone has collapsed. In the southern lowlands of Petén, surface elevation varies between 100 and 300 m above mean sea level and groundwater is fairly inaccessible. The karst landforms of Petén are less developed, however, and surface waters are perched, resulting in numerous lakes and seasonally inundated topographic depressions (Deevey et al. 1979). The lake district contains a series of terminal basins distributed along a series of east-west aligned faults near 17°N latitude. Principal water bodies of the lake chain extend approximately 100 km from westernmost Lake Perdida eastward to the twin basins Yaxhá and Sacnab (Figure 2-1b).

Annual rainfall across the Yucatán Peninsula is highly variable, ranging from a low of 500 mm yr^{-1} along the northwest coasts of Yucatán and Campeche to over 2000 mm yr^{-1} in the southern lowlands (Wilson 1980). Rainfall in Petén varies spatially and inter-annually from ~900 to 2500 mm with a regional mean of 1600 mm (Deevey 1980). Heavy rains between June and October are associated with the northward migration of the inter-tropical convergence zone (ITCZ) and the Azores-Bermuda high-pressure system. This period is characterized by weak trade winds and warm sea surface temperatures in the Atlantic between about 10° and 20°N. Conversely, dry conditions develop in November and December as the ITCZ and the Azores-Bermuda high move equatorward and strong trade winds become predominant (Hastenrath 1976; 1984).

Seasonal aridity and the precipitation gradient in Yucatán are reflected by regional soil development and vegetation distribution. Soils in the northern lowlands are thin, and

support dry-adapted scrub vegetation and some semi-evergreen forest of medium height (Wilson 1980). Farther south, in Petén, the landscape is dominated by well-drained forest soils (Simmons et al. 1959) and tropical semi-deciduous and evergreen vegetation (Lundell 1937). Forest soils of Petén are relatively fertile, but cultivation is restricted by steep slopes (Deevey et al. 1979; Rice et al. 1985). Under deforestation, the erosive potential of intense tropical rain contributes to the rapid transport of dissolved and particulate materials from the land to lakes.

Lake Salpetén

Lake Salpetén (16°58'N and 89°40'W) is a small ($A = 2.6$ km^2), terminal lake basin within the central Petén lake district (Figure 2-1c). The lake lies at 104 m a.s.l. and has a maximum depth of 32 m (Brezonik and Fox 1974) (Figure 2-1c). Lake Salpetén is sulfate-rich (3000 mg l^{-1}) and relatively saline (4500 mg l^{-1} TDS), with a pH of 8.5 and a conductivity of ~3600 μS cm^{-1}. Surface water temperatures typically average between 27° and 30°C throughout the year. Salpetén is an exceptional site for paleoenvironmental research. Evaporative loss from the lake is substantial, as demonstrated by the ~7‰ enrichment of lake water (+4.1‰, $n = 4$) relative to precipitation (-2.7‰, $n = 11$) and groundwater (-3.4‰, $n = 2$). Temporal changes in the $^{18}O/^{16}O$ ratio of the lake water are controlled primarily by changes in the ratio of evaporation to inputs, and past changes in the $^{18}O/^{16}O$ ratio of the lake water are recorded in the $\delta^{18}O$ record of carbonate in gastropod and ostracode shells preserved in the lake sediments. The basin lacks outflows, and sediments are therefore the ultimate sink of most dissolved and particulate matter that enters the lake.

Methods

In May 1980, fifteen meters of sediment were obtained from the deep basin of Lake Salpetén (Figure 2-1d) (Deevey et al. 1983). Sediments were collected with a 45.7-cm split-spoon sampler. The split-spoon sampler did not recover unconsolidated surface sediments, and uppermost retrieved deposits lie approximately 1.6 m below the sediment surface. In July 1997, additional sediment was recovered from Lake Salpetén in a water depth of 9.2 m (SP4-16-VII-97). Surface sediments were collected with a piston corer designed to retrieve undisturbed sediment-water interface profiles (Fisher et al. 1992) and deeper sections were taken in 1.0-m segments with a square-rod piston corer (Wright et al. 1984). The interface core was sectioned in the field at 1.0-cm intervals by upward extrusion into a sampling tray fitted to the top of the core barrel. Square-rod core sections were extruded in the field, stored in PVC pipe, transported to the laboratory, and sampled at 1.0-cm intervals. Archived core sections from the deep basin of Lake Salpetén, designated Sal 80-1, were reexamined in May 2000 and sampled at 5.0-cm intervals to a depth of ~1090 cm.

Sediment ages from Lake Salpetén were determined by accelerator mass spectrometry (AMS) of ^{14}C in terrestrial organic matter (wood, charcoal, and seed). Radiocarbon samples were measured at Lawrence Livermore National Laboratories and the National Ocean Science AMS Facility at Woods Hole Oceanographic Institution. Calibrated dates and calendar ages were calculated using the INTCAL98 program with a 100 year moving average of the tree-ring calibration data set (Stuiver et al. 1998).

Oxygen and carbon isotopic ratios were measured on gastropod (*Cochliopina* sp.) and ostracode shells (*Heterocypris* sp., *Limnocythere* sp., and *Physocypria globula*.). Sediment samples were disaggregated in 3% H_2O_2 and washed through a 63-µm sieve.

Coarse material (>63 μm) was dried at 60°C. Adult ostracode valves and gastropod shells were picked from the dried samples, soaked in 15% H_2O_2, cleaned ultrasonically in deionized water, and rinsed with methanol before drying. Aggregate samples of ~25 ostracode carapaces were measured from each 1.0-cm sediment sample. Gastropod shells of ~15 individuals were ground to a fine powder and a fraction of the ground carbonate was analyzed from each sample.

Carbonate samples for stable isotope analysis from Sal 80-1 were reacted in 100% orthophosphoric acid at 70°C using a Finnigan MAT Kiel III automated preparation system. Isotopic ratios of purified CO_2 gas were measured on-line with a Finnigan MAT 252 mass spectrometer and compared to an internal gas standard. Carbonate samples from SP4-16-VII-97 were reacted in a common acid bath of 100% orthophosphoric acid at 90°C using a VG/Micromass Isocarb preparation system. Stable isotope analysis of the resulting CO_2 gas was determined on-line with a triple-collector VG/Micromass Prism Series II mass spectrometer. All carbonate isotopic values are expressed in conventional delta (δ) notation as the per mil (‰) deviation from Vienna Pee Dee Belemnite (VPDB). Precision for $\delta^{18}O$ and $\delta^{13}C$ of the samples analyzed with the Finnigan MAT 252 was ± 0.09‰ and ± 0.03‰, respectively. Precision of the samples analyzed with the VG/Micromass Prism was ± 0.08‰ ($\delta^{18}O$) and ± 0.06‰ ($\delta^{13}C$).

Trace element (Mg and Sr) concentrations of the ostracodes *Limnocythere* sp. and *Cytheridella ilosvayi* were determined by atomic absorption techniques with a Perkin-Elmer 3100 EDS spectrophotometer and coupled HGA-600 graphite furnace with autosampler. Multiple-valve samples (weighing ~50 μg) were measured after dissolution

in 5.0-mL of 2% nitric acid. Elemental concentrations are expressed in μg g^{-1} by weight of calcium carbonate ($CaCO_3$).

Inorganic carbon (IC) in the sediments was measured by coulometric titration (Engelmann et al. 1985) with a UIC/Coulometrics Model 5011 coulometer and coupled UIC 5240-TIC carbonate autosampler. Analytical precision is estimated to be ± 0.6% based upon repeated analysis of reagent-grade calcium carbonate. Total carbon (TC) in the sediments was measured with a Carlo Erba NA 1500 CNS elemental analyzer with autosampler. Organic carbon (OC) was estimated by subtraction of IC from TC.

Table 2-1. AMS radiocarbon dates for samples from Lake Salpetén, Petén, Guatemala.

Core	Sample Type	Depth (cm)	Radiocarbon Age (yrs B.P.)	Calibrated Age (A.D./B.C.)	Error (± yrs B.P.)
Sal 80-1	Charcoal	157	1820	200 A.D.	190
	Charcoal	395	2200	220 B.C.	70
	Charcoal	503	2200	220 B.C.	60
	Charcoal	646	2430	460 B.C.	50
	Charcoal	707	2500	750 B.C.	60
	Charcoal	774	2990	1240 B.C.	190
	Charcoal	813	3160	1420 B.C.	80
	Charcoal	960	6990	5850 B.C.	50
	Charcoal	1012	8220	7240 B.C.	50
SP4-16-VII-97	Wood	20	480	1440 A.D.	65
	Wood	75	1130	920 A.D.	40
	Wood	97	4130	2650 B.C.	55
	Wood	103	5240	4030 B.C.	60
	Wood	151	8660	7610 B.C.	65
	Wood	162	8670	7620 B.C.	95
	Wood	165	8770	7890 B.C.	65
	Wood	172	8780	7900 B.C.	75

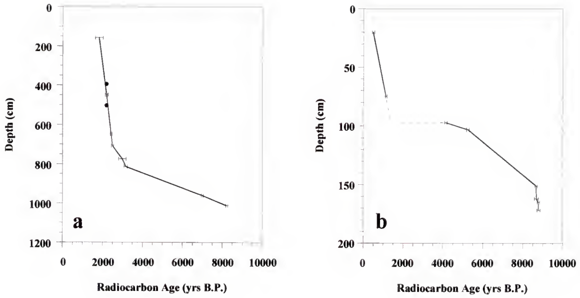

Figure 2-2. Radiocarbon age versus depth for Lake Salpetén sediment cores. Radiocarbon age versus depth for core Sal 80-1 (a) and core SP4-16-VII-97 (b). Age-depth values were determined by linear interpolation between dated horizons. The age-depth profile demonstrates abrupt change in radiocarbon activity after 4130 ^{14}C yrs B.P.

Results

Nine AMS ^{14}C dates were obtained from core Sal 80-1 and yielded a maximum age of 8220 ^{14}C yrs B.P. (Table 2-1). Age-depth values were determined by linear interpolation between dated horizons, and linear sedimentation rates were extrapolated to the base of the sediment profile (Figure 2-2). In core SP4-16-VII-97, eight AMS ^{14}C samples provided a basal age of 8780 ^{14}C yrs B.P. The age-depth profile in the core demonstrates abrupt change in radiocarbon activity after 4130 ^{14}C yrs B.P., suggesting erosion or non-deposition. Following the apparent hiatus in core SP4-16-VII-97, continuous sedimentation is evident since at least 1130 ^{14}C yrs B.P.

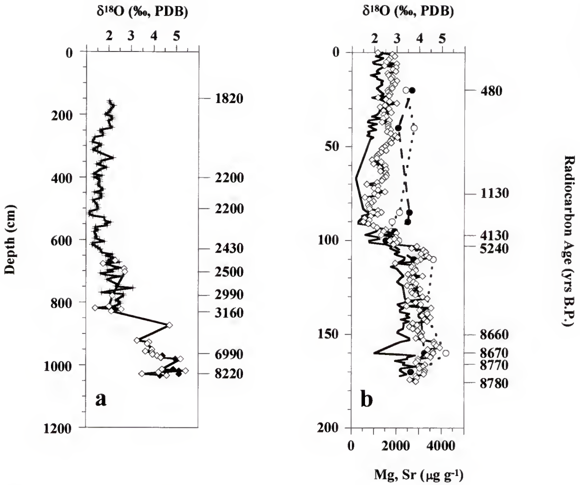

Figure 2-3. Biogenic carbonate $\delta^{18}O$ profiles from Lake Salpetén sediment cores. (a) Oxygen isotopic composition based on the gastropod *Cochliopina* sp. (open diamonds) and two ostracodes, *Physocypria globula* (crosses) and *Heterocypris* sp. (filled diamonds) versus depth in core Sal 80-1. (b) Oxygen isotopic composition based on the gastropod *Cochliopina* sp. (open diamonds) and the ostracode *Limnocythere* sp. (solid line) and Mg (open circles) and Sr (filled circles) concentrations in valves of the ostracod *Limnocythere* sp. versus depth in core SP4-16-VII-97.

Paleoenvironmental proxies from Sal 80-1 and SP4-16-VII-97 are plotted against depth (Figures 2-3 and 2-4) and results are discussed as a function of age (Figure 2-5). The $\delta^{18}O$ of biogenic carbonate in Lake Salpetén sediments was high prior to 7200 ^{14}C yrs B.P. and averaged 4.7‰ (Figure 2-5a). Between 7200 and 3500 ^{14}C yrs B.P., mean

values decreased to ~2.3‰, with the exception of a brief excursion toward higher values centered at 4700 ^{14}C yrs B.P. Similarly, the trace element (Mg and Sr) concentration of biogenic carbonate was relatively high prior to 5240 ^{14}C yrs B.P., but decreased by 4130 ^{14}C yrs B.P. (Figure 2-3b). Magnesium and strontium concentrations remained low after 4130 ^{14}C yrs B.P. Minimum $\delta^{18}O$ values (averaging 1.7‰) occurred between 2400 and 1800 ^{14}C yrs B.P. (Figure 2-5a).

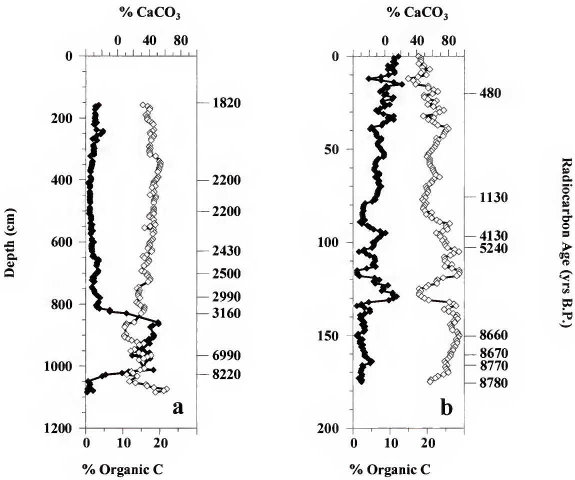

Figure 2-4. Lithologic profiles from Lake Salpetén sediment cores. (a) Organic carbon content (filled diamonds) and $CaCO_3$ concentration (open diamonds) versus depth in core Sal 80-1. (b) Organic carbon content (filled diamonds) and $CaCO_3$ concentration (open diamonds) versus depth in core SP4-16-VII-97.

Basal sediments in Lake Salpetén had relatively high $CaCO_3$ concentrations that decreased to <20% between 9500 and 9000 ^{14}C yrs B.P. (Figure 2-5b). Organic carbon concentration was relatively low (<5%) prior to 9000 ^{14}C yrs B.P., but increased by 8200 ^{14}C yrs B.P. (Figure 2-5c). Organic carbon concentrations remained high (typically >15%) between 8200 and 4300 ^{14}C yrs B.P. Between 9000 and 4800 ^{14}C yrs B.P., $CaCO_3$ content was highly variable, with maxima (30-40%) centered at 8100, 7100, and 6100 ^{14}C yrs B.P. After 4800 ^{14}C yrs B.P., $CaCO_3$ content increased steadily from <30% to ~50%. Organic C concentration decreased from 20% to less than 5% between 4300 and 3200 ^{14}C yrs B.P. and remained low between 3200 and 1800 ^{14}C yrs B.P.

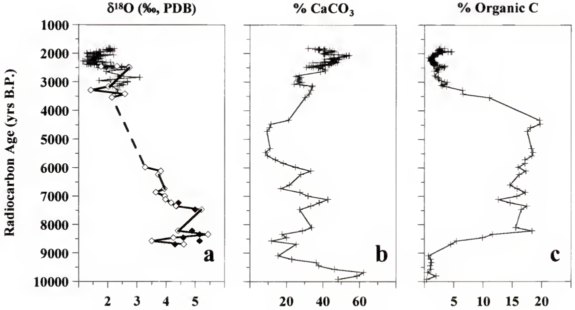

Figure 2-5. Oxygen isotope stratigraphy and lithology of Lake Salpetén sediments versus age. (a) Oxygen isotopic composition based on the gastropod *Cochliopina* sp. (open diamonds) and two ostracods, *Physocypria globula* (crosses) and *Heterocypris* sp. (filled diamonds), (b) $CaCO_3$ concentration, and (c) organic carbon content versus radiocarbon age in Lake Salpetén core Sal 80-1.

Discussion

The oldest date from Salpetén core SP4-16-VII-97 indicates that the lake filled above a depth of 9.2 m below present level by ~9000 ^{14}C yrs B.P. This rise in water level coincided with the filling of Lake Petén Itzá ca. 9000 ^{14}C yrs B.P. (Curtis et al. 1998) and Lake Quexil prior to 8410 ^{14}C yrs B.P. (Vaughan et al. 1985). Early Holocene filling of Petén lakes is attributed to increased moisture availability following the arid late Pleistocene (Leyden 1984; Leyden et al. 1993). Oxygen isotopic values in Lake Salpetén were highest in the early Holocene between ~9000 and 7200 ^{14}C yrs B.P. (Figure 2-6a). The isotope profile from nearby Lake Petén Itzá (Curtis et al. 1998) reveals a similar pattern of variation with the greatest $\delta^{18}O$ values between 9000 and 6800 ^{14}C yrs B.P. (Figure 2-6b). If the $\delta^{18}O$ signal is interpreted as a reflection of E/P, the records suggest that climate was relatively dry during the early Holocene, prior to ~6800 ^{14}C yrs B.P. This interpretation is at odds with palynological evidence from many Petén lakes suggesting moist conditions and extensive lowland forests in the early Holocene from 9000 to 5600 ^{14}C yrs B.P. (Figures 2-6c and 2-6d) (Leyden 1984; 1987; Islebe et al. 1996).

Early Holocene Climate, Vegetation, and Watershed Interactions

The discrepancy between early Holocene climatic inferences from pollen and $\delta^{18}O$ records may be reconciled if lake water $\delta^{18}O$ was controlled not only by E/P, but also by varying inputs of surface and groundwater flow to the lake as a result of changing vegetation density in the watershed. Input waters to Petén lakes display relatively light oxygen isotopic values, whether delivered by direct rainfall (-2.7‰) or surface and groundwater flow (-3.4‰). Reduced input of meteoric waters would decrease lake volume, increase the proportion of the hydrologic budget lost to evaporation, and

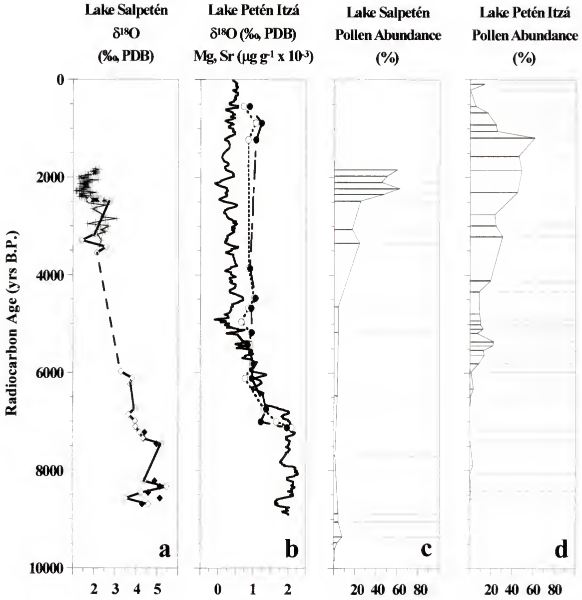

Figure 2-6. Oxygen isotope and pollen records from Lakes Salpetén and Petén Itzá. (a) Oxygen isotopic composition of gastropods and ostracodes from Lake Salpetén core Sal 80-1, (b) $\delta^{18}O$ of gastropods from Lake Petén Itzá (solid line) and Mg and Sr concentrations (open and filled circles, respectively) in valves of the ostracode *Cytheridella ilosvayi*, and relative abundance of pollen types in the (c) Salpetén and (d) Petén Itzá sediment core versus radiocarbon age. Black bars indicate the relative abundance of disturbance taxa and grey bars indicate the relative abundance of lowland forest taxa.

therefore increase lake water $\delta^{18}O$. Dense forest cover in the early Holocene may have increased evapotranspiration and soil moisture storage in the watershed, thereby reducing the catchment water yield and transport of isotopically light surface and ground waters to the lake. As a result, early Holocene $\delta^{18}O$ values of lake water and biogenic carbonate would have been high, explaining the contradictory climatic inferences from pollen and isotopes.

Oxygen isotopic records from Petén lakes may also have been influenced by changes in the isotopic composition of precipitation and/or reduced deepwater temperatures (and consequent high $\delta^{18}O$) during periods of elevated lake level. However, Mg and Sr values in ostracode shells from Lakes Salpetén and Petén Itzá suggest otherwise. Trace element concentrations in Lake Salpetén were highest in the early Holocene between ~9000 and 5200 ^{14}C yrs B.P. (Figure 2-3b). Magnesium and Sr concentrations in Petén Itzá were similarly high prior to ~6800 ^{14}C yrs B.P. (Figure 2-6b). This suggests that the $\delta^{18}O$ records, which parallel trace element concentrations, largely reflect basin hydrology rather than changes in the isotopic composition of precipitation. If the early to mid-Holocene decline in $\delta^{18}O$ reflected an increase in temperature, one would expect to see an increase in Mg values. Covariant decreases in both $\delta^{18}O$ and Mg eliminate temperature as a controlling factor on the lacustrine $\delta^{18}O$ record.

Middle Holocene Deforestation and Hydrologic Change

Middle Holocene pollen records indicate that lowland tropical forest taxa declined as open vegetation taxa increased (Leyden 1984; Vaughan et al. 1985; Leyden 1987; Islebe et al. 1996). For example, lowland forest pollen of the *Moraceae-Urticaceae*

group declined and pollen of the savanna indicator *Byrsonima* increased in the Lake Petén Itzá record as early as ~5600 ^{14}C yrs B.P. This suggests a transition to open forest as a result of climate change (increased E/P) or early human impact (Islebe et al. 1996; Curtis et al. 1998).

The $\delta^{18}O$ values and trace element concentrations of biogenic carbonate in cores from Lakes Salpetén and Petén Itzá decreased, coincident with the decline of lowland forest taxa. The discrepancy between the timing of the decrease in the two lacustrine records probably reflects minor chronological inaccuracies. The $\delta^{18}O$ decrease in Lake Salpetén occurred between ~7200 and 3500 ^{14}C yrs B.P. (Figure 2-6a), whereas $\delta^{18}O$ and Mg and Sr decreased between ~6800 ^{14}C yrs B.P. and 5000 ^{14}C yrs B.P. in Lake Petén Itzá (Figure 2-6b) (Curtis et al. 1998). Pollen and geochemical records in both lakes yield contradictory climate inferences if $\delta^{18}O$ and trace element data are interpreted as reflecting only changes in E/P. Forest decline may have reduced evapotranspiration and soil moisture storage, thereby increasing surface and groundwater flow to the lake basins, resulting in higher lake levels and lower lake water $\delta^{18}O$ values and trace element concentrations.

Late Holocene Climate Change and Human Disturbance

Minimum $\delta^{18}O$ values in Lake Salpetén, suggesting high lake levels, were reached between 3500 and 1800 ^{14}C yrs B.P. (Figure 2-6a) corresponding to the period of Maya occupation. Independent evidence for high water level is provided by twelve AMS ^{14}C dates on aquatic gastropods from subaerial lacustrine deposits ~1.0 to 7.5 m above modern lake stage (Table 2-2). The oldest dates indicate that the lake filled to a depth of more than 7.0 m above modern lake stage between 4600 and 3600 ^{14}C yrs B.P., but most

indicate episodes of high lake levels between 2700 and 1300 ^{14}C yrs B.P. These ages coincide with reduced $\delta^{18}O$ in the core and support the inference for increased water input. In contrast, mean $\delta^{18}O$ values in the Lake Petén Itzá core were nearly constant from 5000 ^{14}C yrs B.P. to present and fluctuated by only ~0.5‰ (Figure 2-6b). The low variability in the late Holocene Petén Itzá record may simply reflect the large volume and long residence time of the lake that make it relatively insensitive to either climatic or land-use changes (Curtis et al. 1998).

High lake levels and $\delta^{18}O$ minima in Lake Salpetén may reflect increased precipitation and reduced evaporation beginning after 3500 ^{14}C yrs B.P. Changes in the $\delta^{18}O$ record may also have been influenced by hydrologic changes in the watershed caused by Maya land-use practices. For example, minimum $\delta^{18}O$ values between 3500 and 1800 ^{14}C yrs B.P. coincide with Maya-induced deforestation around Lakes Salpetén and Petén Itzá (Figures 2-6c and 2-6d; Leyden 1987; Islebe et al. 1996) and other Petén lakes (Tsukada 1966; Deevey et al. 1979; Vaughan et al. 1985). During the same time interval, geochemical records from Lake Salpetén showed changes that reflect catchment soil erosion. Calcium carbonate values in core Sal 80-1 reached a maximum ~2000 ^{14}C yrs B.P. (Figure 2-6b). High carbonate values indicate agricultural forest clearance, accelerated erosion of catchment soils, and transport of soil-derived carbonate to the lake. An increase in inorganic colluvium, termed the "Maya clay", is documented in many Petén lakes during the period of Maya occupation (Cowgill and Hutchinson 1966; Deevey et al. 1979; Brenner 1983; Rice et al. 1985; Binford et al. 1987). As forests were removed by the rapidly expanding Maya population, evapotranspiration within the drainage basins may have been reduced, resulting in increased inflow to the lake.

Table 2-2. AMS radiocarbon dates for gastropod shells (biogenic carbonate) from soil samples (subaerial exposed lake sediments) from the catchment of Lake Salpetén.

Elevation (m)	Depth (cm)	$\delta^{18}O$ (‰, PDB)	$\delta^{13}C$ (‰, PDB)	Radiocarbon Age (yrs B.P.)	Calibrated Age (A.D./B.C.)	Error (± yrs B.P.)
~1.0	0	1.51	-2.39	1510	570 A.D.	50
	10	1.64	-2.29	1540	540 A.D.	40
	20	1.18	-3.18	2150	180 B.C.	40
	30	1.07	-2.55	2290	380 B.C.	40
~4.5	0	1.87	-1.85	1300	690 A.D.	40
	10	1.99	-1.89	1820	200 A.D.	40
	20	1.61	-1.83	2220	220 B.C.	40
~7.5	10	1.08	-3.20	1810	210 A.D.	40
	20	1.05	-3.07	2570	780 B.C.	40
	30	0.83	-4.88	2690	840 B.C.	40
	40	0.77	-2.49	3570	1910 B.C.	40
	50	1.17	-2.13	4570	3350 B.C.	40

Regional Correlations and Mechanisms of Holocene Climate Change

Most pollen records from circum-Caribbean lake sediment cores indicate that the wettest period throughout the lowland Northern Hemisphere tropics occurred between ~7000 and 5000 ^{14}C yr BP (Bradbury et al. 1981; Deevey et al. 1983; Leyden 1985; Piperno et al. 1990; Islebe et al. 1996). Hodell et al. (1991) suggested that this period of high precipitation (low E/P) was related to an increase in the intensity of the annual cycle driven by the Earth's precessional cycle. This explanation is supported by meteorological studies of Caribbean climate that demonstrate a strong correlation between precipitation anomalies and intensity of the annual cycle (Hastenrath 1984). Years with abundant rainfall coincide with an enhancement of the annual cycle when the ITCZ moves farther north during the summer rainy season and farther south during the

winter dry season. Summer perihelion in the Northern Hemisphere, which occurred ~8000 ^{14}C yrs B.P., would have been associated with increased intensity of the annual cycle.

Early and middle Holocene $\delta^{18}O$ records from Petén lakes and Lakes Chichancanab (Mexico) and Miragoane (Haiti) exhibit relatively consistent patterns of millennial-scale change, suggesting that that shifts in E/P were controlled (at least in part) by orbitally forced variations in seasonal insolation that modified the intensity of the annual cycle (Figure 2-7). Direct orbital forcing alone, however, does not fully explain the timing, magnitude, or rate of water level change recorded in the lakes. For example, inferred aridity during the earliest Holocene is more pronounced than predicted by insolation forcing. In Yucatán and Haiti, the early to middle Holocene moist period apparently began considerably earlier than in northern Guatemala. Discrepancies among sites may be a consequence of inaccuracies in age/depth relations. Alternatively, hydrologic processes that governed lake filling may have varied among sites, thereby confounding interpretation of the $\delta^{18}O$ records. Dense forest cover in the lowlands of northern Guatemala may have decreased surface runoff and groundwater flow to the lake basins more effectively than the scrub and open forest vegetation of the Chichancanab and Miragoane catchments (Leyden et al. 1998; Higuera-Gundy et al. 1999). Furthermore, rising sea level probably affected early Holocene filling of low-elevation Lakes Chichancanab and Miragoane, but was not a factor in the filling of inland, higher-elevation Lakes Salpetén and Petén Itzá (~110 m a.s.l.).

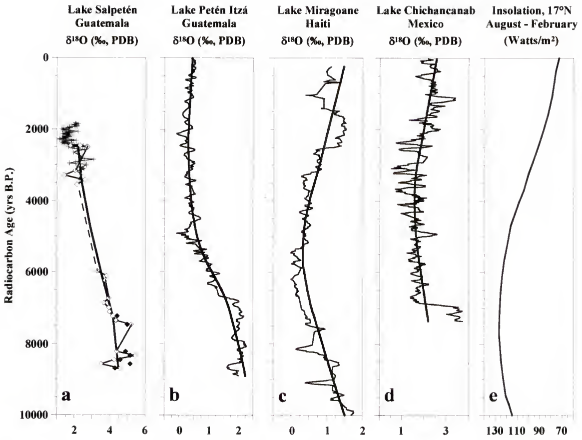

Figure 2-7. Circum-Caribbean $\delta^{18}O$ records and Holocene insolation values. Oxygen isotope records from (a) Lake Salpetén core Sal 80-1, (b) Lake Petén Itzá, (c) Lake Chichancanab, and (d) Lake Miragoane versus radiocarbon age, and (e) the difference in insolation (W m^{-2}) at 17°N latitude between August and February, which is a measure of the intensity of the annual cycle (Hastenrath 1984).

Oxygen isotopic values in Lakes Chichancanab and Miragoane increased abruptly after 3200 ^{14}C yrs B.P., suggesting late Holocene climatic drying (Figures 2-7c and 2-7d). During the same time interval, Lake Miragoane pollen records indicate the decline of mature forests dominated by *Moraceae* and the prevalence of dry forest taxa (Higuera-Gundy et al. 1999). In contrast, $\delta^{18}O$ values in the Petén Itzá core remained relatively invariant after ~5000 ^{14}C yrs B.P. (Figure 2-7b) and minimum $\delta^{18}O$ values between 3500

and 1800 ^{14}C yrs B.P. in Lake Salpetén (Figure 2-7a) indicate high lake level, perhaps the result of altered basin hydrology and increased surface runoff and groundwater inflow. Moreover, late Holocene pollen-based climate reconstructions from Petén lakes have been confounded by the effects of Maya-induced deforestation (Deevey 1978; Deevey et al. 1979; Vaughan et al. 1985; Leyden 1987; Islebe et al. 1996). Neither Lake Chichancanab nor Lake Miragoane is believed to have suffered significant catchment disturbance by human populations prior to ~1400 ^{14}C yrs B.P. (Binford et al. 1987; Leyden et al. 1998). In the absence of human impact, the late Holocene lacustrine $\delta^{18}O$ records from northern Yucatán and Haiti probably reflect climatic variations more accurately. Removal of dry forest, scrub vegetation surrounding Lakes Chichancanab and Miragoane would only slightly increase water yield relative to the increase associated with clearing of lowland forest (Sahin et al. 1996). The best late Holocene paleoclimate records are thus likely to be found in lakes from minimally impacted drainage basins.

Conclusions

Discrepancies between Holocene climatic inferences based on pollen and geochemical records from Petén lakes can be reconciled if lake water $\delta^{18}O$ and trace element concentration was not controlled predominantly by E/P, but rather by the changing input of surface and groundwater to the lakes. Dense early Holocene forests caused high evapotranspiration and soil moisture storage in the watershed, thereby reducing the input of meteoric waters to the lakes. Forest decline after ~5600 ^{14}C yrs B.P. reduced evapotranspiration and soil moisture storage, thereby increasing run-off and groundwater flow to the lakes. High input of meteoric waters to the lakes is reflected by low $\delta^{18}O$ and Mg and Sr values in Lake Salpetén between 7200 and 3500 ^{14}C yrs B.P.

and in Lake Petén Itzá between 6800 ^{14}C yrs B.P. and 5000 ^{14}C yrs B.P. After 3500 ^{14}C yrs B.P., $\delta^{18}O$ and trace element values in Lake Salpetén decreased, reaching minimum values between 3500 and 1800 ^{14}C yrs B.P. High lake levels were confirmed by AMS ^{14}C dates of aquatic gastropods from subaerial soil (i.e., sediment) profiles that yielded ages ranging from 4570 to 1510 ^{14}C yrs B.P. High lake levels may have resulted from increased precipitation (reduced E/P) and/or increased surface runoff and groundwater inflow as a consequence of human-induced deforestation. Results from Petén lakes suggest that natural or anthropogenic changes in vegetation cover within watersheds can alter lake hydrologic budgets, thereby confounding paleoclimatic inferences based on the $\delta^{18}O$ and trace element composition of biogenic carbonate.

CHAPTER 3
A 4000-YEAR RECORD OF ENVIRONMENTAL CHANGE FROM THE SOUTHERN MAYA LOWLANDS, PETEN, GUATEMALA

Introduction

Pollen records from lowland Petén, Guatemala, document forest clearance by Maya populations beginning in the first millennium B.C. (Tsukada 1966; Deevey 1978; Vaughan et al. 1985; Leyden 1987; Islebe et al. 1996). During the later Preclassic and Classic Periods (~300 B.C. to 850 A.D.), as Maya settlement and agriculture expanded, much of Petén was deforested. Widespread cultivation accelerated soil erosion, which resulted in the accumulation of thick clay-rich deposits in many Petén lakes (Cowgill and Hutchinson 1966; Deevey et al. 1979; Brenner 1983; Rice et al. 1985; Binford et al. 1987). This inorganic sediment accumulation, termed the Maya clay, was deposited over a period of ~2600 years and is equated with occupation of Petén watersheds (Binford et al. 1987). Forest recovery and soil stabilization has been attributed to population decline following the Classic Maya Period and further depopulation after European contact (Brenner et al. 1990).

The chronological framework of Maya impact on terrestrial and lacustrine environments remains imprecise. Hard-water lake effect and re-deposition of carbonate-rich basin soils of unknown isotopic composition (Deevey and Stuiver 1964) invalidate radiocarbon dates on bulk sediments from Petén lakes. Chronologies have therefore relied largely on correlation between pollen stratigraphies and the archaeological prehistory of the region (Vaughan et al. 1985; Brenner 1994). Moreover, late Holocene

climate reconstructions from pollen analyses are confounded by human-mediated deforestation (Deevey 1978; Vaughan et al. 1985; Leyden 1987; Islebe et al. 1996).

In this study, a composite sediment profile from Lake Salpetén provides the first high-resolution, accelerator mass spectrometry (AMS) ^{14}C-dated record of late Holocene environmental change in lowland Petén, Guatemala. Stratigraphic geochemical variations from this profile reflect changing hydrologic balance and material transfer to the lake resulting from both human disturbance and climate change. Shifts in the water balance of the lake were inferred from the oxygen isotopic composition ($\delta^{18}O$) of biogenic carbonate. Material transfer to the lake was inferred from sediment composition and accumulation, and watershed vegetation changes were reconstructed from fossil pollen assemblages. The Lake Salpetén record was compared with a profile from nearby Lake Petén Itzá (Curtis et al. 1998) and sediment records from the northern Yucatán Peninsula (Hodell et al. 1995; Curtis et al. 1996).

Study Site

The Department of Petén occupies the northern third of Guatemala (Figure 3-1a) and is characterized principally by low-lying karsted limestones of Cretaceous and Tertiary age (Vinson 1962). The landscape is dominated by well-drained forest soils and tropical semi-deciduous and evergreen vegetation (Lundell 1937). Terrain varies between 100 and 300 m above mean sea level. Groundwater lies well below the land surface and is fairly inaccessible. Surface waters, however, are perched resulting in numerous lakes and seasonally inundated topographic depressions (Deevey et al. 1979). The lake district contains a series of terminal basins along a series of east-west aligned *en echelon* faults at 17°N latitude (Figure 3-1b). Lake Salpetén is a small closed-basin lake

that lies at 104 m a.s.l. and has a maximum depth of 32 m (Brezonik and Fox 1974) (Figure 3-1c). Lake Salpetén is sulfate-rich (3000 mg l^{-1}) and relatively saline (4500 mg l^{-1} TDS), with a pH of 8.5 and a conductivity of ~3600 $\mu S\ cm^{-1}$.

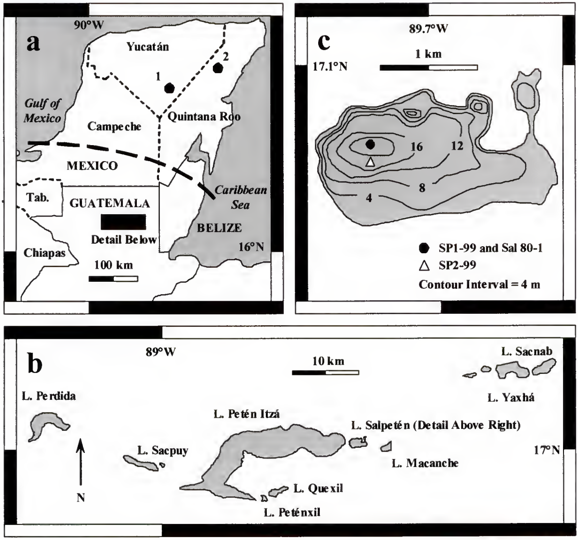

Figure 3-1. Lake study sites in the Yucatán Peninsula and Department of Petén, Guatemala. (a) Map showing lake study sites in northern Guatemala (black rectangle) and Lakes Chichancanab (1) and Punta Laguna (2). Heavy dashed line indicates the division between the northern and southern Maya lowlands. (b) Petén lake district in Guatemala. (c) Bathymetry and sediment core sites in Lake Salpetén.

Rainfall in Petén varies spatially and inter-annually from 900 to 2500 mm, with a regional annual mean of 1600 mm (Deevey 1978). Intense convection associated with the northward migration of the inter-tropical convergence zone (ITCZ) and the Azores-Bermuda high-pressure system produces heavy rains between June and October when the trade winds weaken and sea-surface temperatures warm in the Atlantic between 10° and 20°N (Hastenrath 1984). Conversely, dry conditions develop in November and December as the ITCZ and the Azores-Bermuda high move equatorward and strong trade winds become predominant. A pronounced dry season prevails from January to May.

Materials and Methods

In May 1980, fifteen meters of sediment were obtained from the deep basin of Lake Salpetén (Deevey et al. 1983) with a 45.7-cm split-spoon sampler. This core was designated Sal 80-1. Topmost sediments, above 1.6 m, were not recovered. In August 1999, two additional sediment cores were recovered from Lake Salpetén in water depths of 16.3 (SP2-19-VIII-99) and 23.2 m (SP1-17-VIII-99) (Figure 3-1c). Surface sediments were collected with a sediment-water interface corer (Fisher et al. 1992). Deeper sections were taken in 1.0-m segments with a modified square-rod piston corer (Wright et al. 1984). Interface cores were sectioned in the field at 1.0-cm intervals. Square-rod core sections were extruded and sampled at 1.0-cm intervals in the laboratory. Archived core sections from Sal 80-1 were sampled at 5.0-cm intervals and reexamined in May 2000.

Sediment ages from Lake Salpetén were determined by accelerator mass spectrometry (AMS) of ^{14}C in terrestrial organic matter (wood, seeds, and charcoal) at Lawrence Livermore National Laboratories. Calibrated dates and calendar ages were

calculated using the INTCAL98 calibration with a 100-year moving average of the tree-ring calibration data set (Stuiver et al. 1998).

Oxygen isotopic ratios were measured on valves of the ostracode *Physocypria globula*. Adult ostracode valves were soaked in 15% H_2O_2, cleaned ultrasonically in deionized water, and rinsed with methanol before drying. Aggregate samples of ~40 ostracode carapaces were measured from each sediment sample. Samples were reacted in 100% orthophosphoric acid at 70°C using a Finnigan MAT Kiel III automated preparation system. Isotopic ratios of purified CO_2 gas were measured on-line with a Finnigan MAT 252 mass spectrometer and compared to an internal gas standard. Isotopic values are expressed in conventional delta (δ) notation as the per mil (‰) deviation from Vienna Pee Dee Belemnite. Precision for $\delta^{18}O$ samples was ± 0.09‰.

Inorganic carbon (IC) was measured by coulometric titration (Engelmann et al. 1985) with a UIC/Coulometrics Model 5011 coulometer and coupled UIC 5240-TIC carbonate autosampler. Analytical precision is about ± 0.6% based upon repeated analysis of reagent-grade calcium carbonate. Total carbon (TC) was measured with a Carlo Erba NA 1500 CNS elemental analyzer with autosampler. Organic carbon (OC) was estimated by subtraction of IC from TC.

Results

The high degree of stratigraphic correlation between Lake Salpetén sediment cores allowed construction of a composite depth series comprised of seventeen AMS ^{14}C dates (Table 3-1). Age-depth values for the last 4000 calibrated radiocarbon years were fit by a fourth-order polynomial (Figure 3-2). Residual errors associated with the construction of the composite depth series averaged ± 110 years.

Table 3-1. AMS radiocarbon dates for samples from Lake Salpetén, Petén, Guatemala.

Core	Sample Type	Depth (cm)	Radiocarbon Age (yrs B.P.)	Calibrated Age (cal yr A.D. or B.C.)	Error (± yrs B.P.)
SP1-99	Wood	24	180	1670 A.D.	50
SP2-99	Wood	59	400	1460 A.D.	40
SP2-99	Wood	108	1370	660 A.D.	50
SP2-99	Charcoal	114	1380	660 A.D.	140
SP1-99	Wood	123	1320	690 A.D.	40
SP2-99	Charcoal	125	1690	370 A.D.	140
SP2-99	Seed	153	1850	170 A.D.	50
Sal 80-1	Charcoal	174	2200	220 B.C.	70
Sal 80-1	Charcoal	221	2200	220 B.C.	60
SP2-99	Wood	231	2090	100 B.C.	40
SP2-99	Seed	236	2200	220 B.C.	50
Sal 80-1	Charcoal	271	2430	460 B.C.	50
Sal 80-1	Charcoal	296	2500	750 B.C.	60
SP2-99	Wood	309	2820	960 B.C.	40
Sal 80-1	Charcoal	323	2990	1240 B.C.	190
SP2-99	Wood	339	3240	1510 B.C.	40
Sal 80-1	Charcoal	344	3160	1420 B.C.	80

Paleoenvironmental proxies from Lake Salpetén are plotted and discussed relative to calendar age (cal yrs A.D. or B.C.). Calcium carbonate concentrations in Lake Salpetén sediments increased gradually from 2000 cal yrs B.C. to maximum values (>40%) between 700 cal yrs B.C. and 850 cal yrs A.D. (Figure 3-3d). Organic carbon concentration was relatively high, typically >10%, before 1700 cal yrs B.C., but decreased by 900 cal yrs B.C. (Figure 3-3c). Organic carbon concentrations remained low (<5%) between 900 cal yrs B.C. and 850 cal yrs A.D. Organic carbon concentration increased abruptly after 850 cal yrs A.D. and again after 1400 cal yrs A.D., but declined over the last 300 years. After 850 cal yrs A.D., $CaCO_3$ content was highly variable.

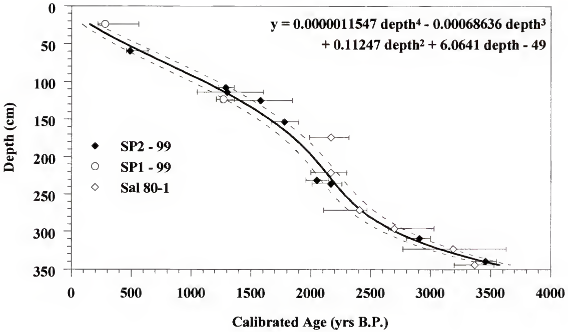

Figure 3-2. Sediment age versus composite depth in Lake Salpetén. Age-depth values were fit by a fourth-order polynomial (black line). Error bars delineate calibrated age range at two standard deviations. Dashed lines indicate ±110 year error envelope. Residual error envelopes propagate errors in analysis and calibration.

The $\delta^{18}O$ of biogenic carbonate decreased between 1300 and 400 cal yrs B.C., from ~2.7‰ to less than 1.5‰ (Figure 3-3e). Minimum $\delta^{18}O$ values, averaging 1.3‰, occurred between 400 cal yrs B.C. and 150 cal yrs A.D. Oxygen isotopic values increased abruptly between 150 and 200 cal yrs A.D., and averaged ~1.7‰ between 200 and 500 cal yrs A.D. Between 500 and 550 cal yrs A.D., $\delta^{18}O$ increased by as much as 1.5‰ and averaged 2.2‰ between 550 and 850 cal yrs A.D. Average $\delta^{18}O$ increased to 2.4‰ after 900 cal yrs A.D., and increased to 2.8‰ between 1300 cal yrs A.D. and the present.

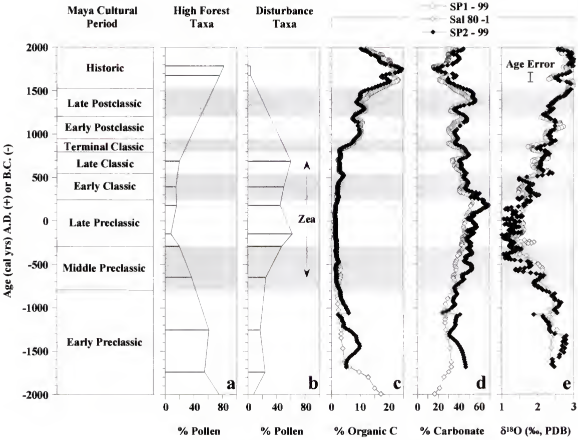

Figure 3-3. Oxygen isotope, sediment lithology, and pollen records from Lake Salpetén. (a) Relative abundance of high forest pollen taxa and (b) pollen taxa typical of disturbed lands from Lake Salpetén core Sal 80-1 (Leyden 1987). High forest taxa include species of the Moraceae (e.g., Brosimum, Cecropia, Chlorophora, and Ficus) and Urticaceae families. Disturbance taxa include Amaranthaceae, Ambrosia, Compositae, Cyperaceae, and Gramineae. (c) Organic carbon content, (d) $CaCO_3$ concentration, and (e) oxygen isotopic composition versus age and Maya cultural periods (Rice and Rice 1990). Geochemical data were smoothed with a 5-point running mean to illustrate long-term trends. Error bar delineates the average residual age error (Figure 3-2).

Interpretation of Proxy Records

Variations in the oxygen isotopic ratio ($^{18}O/^{16}O$) of lacustrine carbonate are caused by changes in the temperature of carbonate precipitation and shifts in the $^{18}O/^{16}O$ of lake water from which the carbonate precipitates (Craig 1965). In tropical lakes with negligible surface outflow, variations in the $^{18}O/^{16}O$ of lake water are dominated by

changes in the rate of evaporation relative to the combined inputs of precipitation and surface and groundwater inflow (Fontes and Gonfiantini 1967; Talbot 1990; Curtis et al. 1996; Rosenmeier et al. 2002b). Extended periods of enhanced evaporation or reduced precipitation and surface and groundwater inflow result in higher $^{18}O/^{16}O$ ratios of lake water and precipitated carbonate. Conversely, increased surface and groundwater inflow or precipitation results in lower $^{18}O/^{16}O$ ratios.

In addition to altering lake hydrology, changes in surface flow and precipitation alter the transfer of dissolved and particulate material to a water body. Changes in material flux to the sediments result primarily from variations in material output from the catchment, sometimes due to land-use changes. For example, soil changes resulting from deforestation and fires produce variations in the flux of particulate matter to a lake (Walling 1988; Dearing 1991). Forest clearance and high rainfall accelerate alluviation and colluviation, and increases in sediment accumulation may reflect intensified surface flow and erosion within a watershed. Catchment deforestation, agricultural production, field abandonment, and subsequent plant succession are reflected by pollen deposited in lake sediments. Deforestation alters the transport of soil nutrients and organic and inorganic matter to the lake, which is reflected in the sediment lithologic composition (Binford et al. 1987).

Discussion

Human occupation and deforestation of the Lake Salpetén watershed is documented by reduction of high forest taxa, particularly pollen of the arboreal family Moraceae, and expansion of disturbance taxa beginning ~1700 cal yrs B.C. (Figures 3-3a and 3-3b). During the same time interval, geochemical records from Lake Salpetén reflect catchment soil erosion. Organic carbon content decreased after 1700 cal yrs B.C. and inorganic

materials dominated subsequent sediment accumulation (Figures 3-3c and 3-3d). Forest clearance by the rapidly expanding Maya population destabilized watershed soils, thereby accelerating their erosion, transport, and re-deposition. Accelerated erosion and oxidation of exposed catchment soils rapidly depleted organic matter (forest litter and surface soil horizons) and enhanced delivery of soil-derived carbonate to the lake, resulting in the deposition of sediments with low organic content.

Archaeologically documented colonization of the Salpetén watershed dates to Middle Preclassic times, beginning 1000 B.C. (Rice and Rice 1990). Rapid accumulation of inorganic sediments in Lake Salpetén began before dense occupation of the basin in the Late Classic (600 to 800 A.D.). Catchment response to vegetation removal may not have been linearly related to population expansion and initial forest clearance and soil destabilization may be attributed to pioneer settlement that is archaeologically unrecognized. Intensified basin occupation and deposition of the Maya clay correlates with increased disturbance taxa and the presence of maize (*Zea*) after 700 cal yrs B.C. (Figure 3-3b). Gradual decline of lowland forest and subsequent soil destabilization might also reflect climatic drying prior to human disturbance. However, forest loss as a result of increased aridity is incompatible with the coincident decrease in the $\delta^{18}O$ of biogenic carbonate in Lake Salpetén, beginning ~1300 cal yrs B.C. (Figure 3-3e).

The $\delta^{18}O$ values in Lake Salpetén decreased gradually from 1300 to 400 cal yrs B.C. (Figure 3-3e). Minimum $\delta^{18}O$ values, suggesting high water levels, occurred between 400 cal yrs B.C. and 150 cal yrs A.D., corresponding to the Middle and Late Preclassic Maya Periods. Evidence for high lake levels is also provided by twelve AMS ^{14}C dates on aquatic gastropods from soil pits (i.e., subaerial lacustrine deposits) ~1.0 to

7.5 m above modern lake stage (Rosenmeier et al. 2002b). Several dates precede human disturbance of the watershed but most indicate high lake levels during Maya occupation, from the Middle Preclassic through the late Classic. These ages coincide roughly with $\delta^{18}O$ minima in the core and support the inference for high lake stage. Forest removal may have decreased evapotranspiration and soil moisture storage in the watershed, thereby increasing catchment water yield (Bosch and Hewlett 1982; Stednick 1996) and transport of isotopically light surface and ground waters to the lake. Increased delivery of meteoric waters would increase lake volume, decrease the proportion of the hydrologic budget lost to evaporation, and decrease lake water $\delta^{18}O$. Alternatively, high water levels and $\delta^{18}O$ minima may reflect increased precipitation and reduced evaporation beginning after 1300 cal yrs B.C.

In contrast, mean $\delta^{18}O$ values from nearby Lake Petén Itzá have been nearly constant over the last ~4000 years and fluctuated by only ~0.5‰ (Figure 3-4b). Low variability in the late Holocene Petén Itzá record may simply reflect the large volume and long residence time of the lake that make it relatively insensitive to either climatic or land-use changes. Moreover, the Maya clay is not documented as a distinct stratigraphic unit in the Petén Itzá core (Curtis et al. 1998) although accelerated forest clearance is palynologically documented by 1000 cal yrs B.C. (Islebe et al. 1996). Lake Petén Itzá is substantially larger than Salpetén and the basin may be effectively buffered from watershed disturbances. Furthermore, erosional contributions to the coring site were probably low, as the site lies nearly 1 km from the steep north shore (Curtis et al. 1998).

Oxygen isotopic records from Petén lakes may also have been influenced by changes in the isotopic composition of precipitation and/or temperature changes. Trace

element (Mg and Sr) concentrations in ostracode shells from Lakes Salpetén and Petén Itzá parallel the $\delta^{18}O$ records, suggesting that $\delta^{18}O$ records reflect basin hydrology rather than changes in the isotopic composition of rainfall (Rosenmeier et al. 2002b). Furthermore, covariance in both $\delta^{18}O$ and Mg eliminates temperature as a control on the oxygen isotope record.

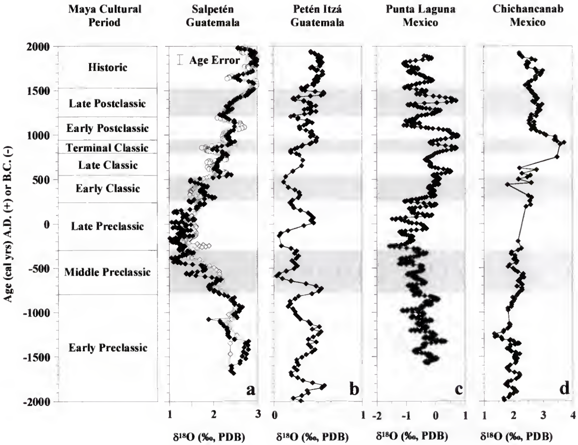

Figure 3-4. Stable isotope records from the Yucatán Peninsula. Oxygen isotope records from (a) Lake Salpetén, (b) Lake Petén Itzá (Curtis et al. 1998), (c) Punta Laguna (Curtis et al. 1996), and (d) Lake Chichancanab (Hodell et al. 1995) versus age and Maya cultural periods (Rice and Rice 1990). Error bar delineates the average residual age error.

Reduced soil erosion in the Salpetén watershed after 850 cal yrs A.D. is inferred from increased organic carbon deposition that coincides with palynological evidence of reforestation (Figures 3-3a to 3-3c). This environmental change coincides with the archaeologically documented decline of Maya population between 800 and 900 A.D. (Lowe 1985). Reduced population in Petén watersheds following the Terminal Classic would have permitted recovery of some lowland forest vegetation and soil stabilization (Deevey et al. 1979; Rice et al. 1985; Binford et al. 1987). Postclassic population densities in Petén, however, were probably sufficient to keep the region partially deforested (Brenner 1994). The increase in organic carbon at the Late Postclassic/Historic boundary (Figure 3-3c) delineates the upper temporal boundary of the Maya clay and is attributed to further decline of Maya populations following European contact.

Forest regeneration after 850 cal yrs A.D. may have altered the hydrologic budget of Lake Salpetén by reducing surface and groundwater inflow, thereby causing an increase in $\delta^{18}O$ during the Postclassic and Historic Periods (Figure 3-3e). This interpretation is at odds with evidence for increased $\delta^{18}O$ values after 150 cal yrs A.D., nearly 700 years prior to Terminal Classic depopulation of the catchment and forest recovery. Late Holocene climate changes may have disrupted regional evaporation and precipitation patterns, altering the hydrologic budget of Lake Salpetén. Strict climatic interpretation of the $\delta^{18}O$ record indicates relatively moist conditions during Middle and Late Preclassic settlement expansion, from 400 cal yrs B.C. to 150 cal yrs A.D. (Figure 3-3e). Abrupt $\delta^{18}O$ increases centered at 150, 550, and 850 cal yrs A.D. document stepwise climatic drying throughout the Classic and Early Postclassic. Maximum $\delta^{18}O$ values

after 1300 cal yrs A.D. reflect the driest conditions of the last ~4000 years, during the Late Postclassic and Historic Period.

Temporal Correlations Between Environmental and Cultural Changes

The paleoclimate history inferred from the Lake Salpetén $\delta^{18}O$ record indicates that shifts in moisture availability coincided with several discontinuities in Maya culture. The termination of Preclassic culture at the site of El Mirador, northern Guatemala (Dahlin 1983) correlates temporally with inferred relative aridity between 150 and 200 cal yrs A.D. (Figure 3-3e). Moreover, the $\delta^{18}O$ increase between 500 and 550 cal yrs A.D. corresponds to the boundary between the Early and Late Classic Periods (Figure 3-3e). This "Maya Hiatus" represents a period of social upheaval and localized population declines (Gill 2000) that may have been associated with climatic drying. Inferred aridity between 850 and 900 cal yrs A.D. (Figure 3-3e) occurred concomitant with the Terminal Classic Maya population decline between 800 and 900 A.D. (Lowe 1985). Climate remained relatively dry throughout the Early Postclassic and $\delta^{18}O$ increased further during the Late Postclassic Period.

Rather than reflecting climate changes, shifts in the Salpetén $\delta^{18}O$ record may simply reflect altered basin hydrology (varied surface runoff and groundwater inflow) related to forest clearance. Stepwise $\delta^{18}O$ increases after 150 cal yrs A.D. may reflect periodic relaxation of catchment disturbance, temporary forest regrowth and soil stabilization, and establishment of new lake level and $\delta^{18}O$ steady-state conditions. Increased $\delta^{18}O$ at the Early Classic/Late Classic boundary coincides with a distinct organic carbon peak that may indicate reduced catchment disturbance and consequent soil stabilization.

Comparison with Northern Yucatán Sediment

The changing water level record from Lake Salpetén both contrasts with and complements late Holocene paleoenvironmental histories inferred from northern Yucatán lake cores. Lake Salpetén levels were relatively high during the Middle and Late Preclassic Period (Figure 3-4a). Oxygen isotopic values increased abruptly at the boundary between the Preclassic and Classic, and continued to increase stepwise throughout the Classic Period. Similarly, the record from Lake Punta Laguna (Curtis et al. 1996) displayed relatively low, but variable $\delta^{18}O$ throughout the Preclassic Period (Figure 3-4c). Oxygen isotopic values in Punta Laguna increased abruptly after 250 A.D. and remained high throughout the Classic Period, suggesting dry conditions. In Lake Chichancanab, $\delta^{18}O$ increased abruptly at 800 A.D., marking the beginning of a 200-year drought in the Terminal Classic and earliest Postclassic (Figure 3-4d) (Hodell et al. 1995; 2001). Oxygen isotope values in Salpetén increased throughout the Postclassic and Historic Periods, whereas $\delta^{18}O$ values in Punta Laguna and Chichancanab decreased after 1100 A.D.

Neither the Chichancanab nor Punta Laguna watersheds are believed to have been densely settled by the Maya (Castillo and Peraza 1991; Leyden et al. 1998). Given this minimal human impact, the $\delta^{18}O$ records from the northern Yucatán lakes probably reflect climatic variations (evaporation and precipitation) accurately. Furthermore, differences between the northern and southern lowlands with respect to vegetation stature and annual precipitation also influenced the relative importance of vegetation in controlling lacustrine hydrologic budgets. For example, clearing of scrub vegetation that predominates in the Chichancanab catchment would increase water yield only about 5%

relative to the increase associated with clearing lowland forest (Sahin et al. 1996). Even if the drying events recorded in northern Lakes Chichancanab and Punta Laguna were regional in extent, human-mediated changes in catchment hydrology in Petén may have obscured the climate signal. The effects of Maya deforestation therefore confound paleoclimatic interpretation of late Holocene $\delta^{18}O$ records from Petén lakes.

Conclusions

Palynological and geochemical records from Lake Salpetén indicate Maya-induced forest clearance and consequent soil erosion beginning ~1700 cal yrs B.C. Reduced soil erosion after 850 cal yrs A.D. coincided with the Terminal Classic Maya demographic decline. Forest recovery and increased organic carbon sedimentation after 1400 cal yrs A.D. correlate with further depopulation of the watershed. Decreased $\delta^{18}O$ of biogenic carbonate between 1300 and 400 cal yrs B.C. coincided with palynological evidence of forest loss. Low $\delta^{18}O$ and inferred high lake levels may have resulted from increased surface runoff and groundwater inflow as a consequence of human-induced deforestation and/or increased precipitation and reduced evaporation. Strictly climatic interpretation suggests higher precipitation during the expansion of Middle and Late Preclassic Maya settlement. Alternatively, minimum $\delta^{18}O$ values between 400 cal yrs B.C. and 150 cal yrs A.D. may have been a consequence of increased surface runoff and groundwater inflow to the lake related to Maya deforestation of the watershed. During the period of Preclassic abandonment (150 A.D.) and the Early Classic/Late Classic boundary (550 A.D.) $\delta^{18}O$ values increased as a consequence of decreased precipitation or temporary forest recovery. Similarly, $\delta^{18}O$ values increased coincident with the Terminal Classic Maya demographic decline. The Lake Salpetén record suggests that anthropogenic

deforestation can alter lake hydrologic budgets, thereby confounding paleoclimatic inferences based on the $\delta^{18}O$ of biogenic carbonate.

CHAPTER 4
QUANTITATIVE ASSESSMENTS OF HOLOCENE ENVIRONMENTAL CHANGE IN PETÉN, GUATEMALA: PREDICTIVE MODELS OF CATCHMENT HYDROLOGY AND LAKE WATER $\delta^{18}O$

Introduction

Variations in the oxygen isotopic composition of biogenic carbonate from Lake Salpetén sediment cores have been used to infer past environmental changes in the southern Maya lowlands of the Yucatán Peninsula (Rosenmeier et al. 2002a; 2002b). These sediment variables indicate that pronounced changes in watershed hydrologic balance were caused by human-induced deforestation and natural climate change. Strictly climatic interpretation of the oxygen isotope record suggests higher precipitation during the period of Maya settlement expansion between 2400 and 1800 cal yrs B.P. Alternatively, this period of minimum $\delta^{18}O$ values may have been a consequence of increased surface runoff and groundwater inflow to the lake related to watershed deforestation by the Maya. When Maya population declined beginning ~1100 years ago, $\delta^{18}O$ values increased as a consequence of reduced hydrologic input to the lake caused by decreased precipitation or forest recovery.

In this study, a hydrologic and isotopic mass balance model of Lake Salpetén provides the first quantitative assessment of the impact of climate and catchment vegetation changes on lake water $\delta^{18}O$ in lowland northern Guatemala. Model simulations of lake inflows and outflows and associated $\delta^{18}O$ values were constrained by a set of modern meteorological parameters and known changes in forest cover. Experiments also tested the plausibility of systematic changes in late Holocene

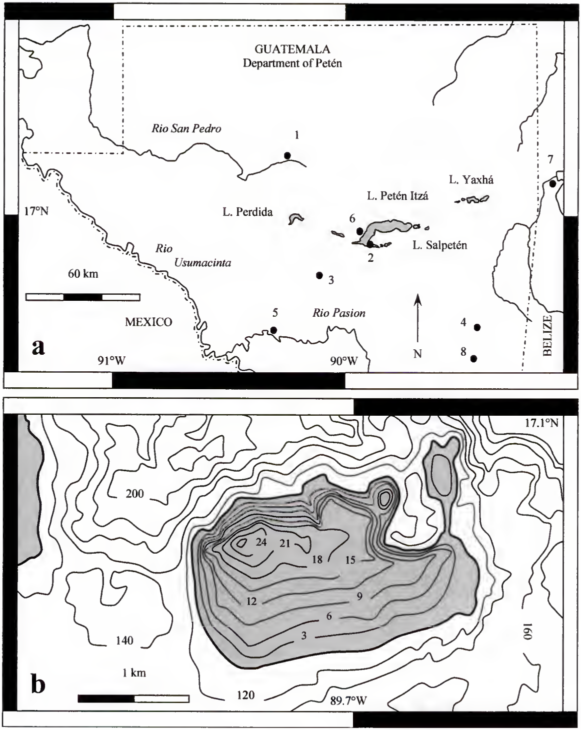

Figure 4-1. The Department of Petén, northern Guatemala and Lake Salpetén. (a) Map showing lake study sites and meteorological stations discussed in the text: (1) El Paso, (2) Flores, (3) Libertad, (4) Poptun, (5) Porvenir, (6) San Andres, (7) San Ignacio (8) San Luis. (b) Bathymetry and topography of Lake Salpetén and the surrounding watershed. Topographic and bathymetric contours appear at intervals of 20 and 3 m, respectively.

precipitation as a control on the hydrologic and isotopic evolution of the lake. These model results were compared with measured $\delta^{18}O$ profiles from Lake Salpetén and sediment records from the northern Yucatán Peninsula.

Study Site

The karstic landscape of the Department of Petén, Guatemala, (Figure 4-1a) is characterized principally by well-drained forest soils and tropical semi-deciduous and evergreen vegetation (Lundell 1937). Terrain varies between 100 and 500 m above sea level and groundwater lies well below the land surface. Surface waters, however, are perched, resulting in numerous lakes and seasonally inundated topographic depressions. The lake district contains a number of terminal basins distributed along a series of east-west aligned faults centered at ~17°N latitude. Principal water bodies of the lake chain extend approximately 100 km from westernmost Lake Perdida eastward to the twin basins of Lakes Yaxhá and Sacnab. Lake Salpetén is a small water body (2.55 km^2) with a catchment area of 6.36 km^2 (Figure 4-1b) and no surface inflows or outflows. The lake lies ~104 m above sea level and has a maximum depth of 32 m (Brezonik and Fox 1974).

Rainfall in Petén varies spatially and interannually from ~1500 to 4100 mm (Instituto Nacional de Sismología, Vulcanología, Meteorología, e Hidrología) (Table 4-1). Heavy rains between June and October are associated with northward migration of the inter-tropical convergence zone (ITCZ) and the Azores-Bermuda high-pressure system. Dry conditions develop in November and December as the ITCZ and Azores-Bermuda high move equatorward and strong trade winds become predominant (Hastenrath 1984). Mean monthly air temperatures in Petén vary between ~20 and 29°C although diurnal temperatures demonstrate a much greater range of variability between ~18 and 33°C.

Table 4-1. Climate Statistics for sites within Petén, Guatemala and western Belize.

Location	Elevation (m)	Temperature[h] (°C)	Rainfall (mm yr^{-1})	Humidity (%)	Windspeed (m s^{-1})	Evaporation (mm mo^{-1})
El Paso [a]	115	29.5 / 20.7	1720	–	–	–
Flores [b]	123	31.4 / 20.0	1555	78	–	–
Flores [c]	–	–	1540	–	–	–
Flores [d]	–	31.9 / 21.6	–	77	1.7	–
Flores [e]	–	29.2 / 20.8	1510	–	–	90
Libertad [b]	125	31.3 / 19.7	1845	82	3.0	105
Poptun [b]	500	28.6 / 18.2	1850	82	–	–
Porvenir [f]	125	29.2 / 19.9	1860	–	–	–
San Andres [b]	150	32.8 / 19.7	1630	82	–	115
San Ignacio [g]	127	–	1550	–	–	–
San Luis [b]	190	31.2 / 20.2	4125	81	–	–

[a] Values for the period 1924-1934 (Lundell 1937).
[b] Values for the period 1997-2002 (INSIVUMEH).
[c] Values for the period 1973-1987 (GHCN – Vose et al. 1992).
[d] Values for the period 1998-2003 compiled from daily data at the Flores airport.
[e] Values for the period 1973-1974 (Deevey et al. 1980).
[f] Values for the period 1968-1983 (GHCN – Vose et al. 1992).
[g] Values for the period 1966-1979 (GHCN – Vose et al. 1992).
[h] Average maximum and minimum temperature values.

Model Description and Methods

The hydrologic balance of a lake (∂V_L) is controlled by the transfer of water to and from the catchment according to the equation:

$$\partial V_L = \Sigma I + P - \Sigma O - E \quad (1)$$

where ΣI and ΣO are the total surface and sub-surface inflows to (I) and outflows (O) from the lake, P is direct precipitation over the lake, and E is the evaporative loss from the lake (Dinçer 1968; Gat 1981). A similar expression can be written for the oxygen isotopic composition of lake water (δ_L):

$$\partial V_{\mathrm{L}}\, \delta_{\mathrm{L}} = \Sigma\, I\, \delta_{\mathrm{I}} + P\, \delta_{\mathrm{P}} - \Sigma\, O\, \delta_{\mathrm{O}} - E\, \delta_{\mathrm{E}} \tag{2}$$

where δ is the isotopic composition of the various inputs and outputs. In lakes with negligible surface outflow, the hydrologic balance and lake water isotopic composition is determined by the difference between evaporation and precipitation over the lake, the water balance of the surrounding catchment (ΣI), and losses through sub-surface (downward) leakage (ΣO) (Deevey 1988). These simple mass balance equations represent the governing components of the hydrologic and isotopic mass balance models constructed with *STELLA* (High Performance Systems™) software.

Hydrologic Model Equations

The model calculates lake water balance through the volumetric (m^3) addition of direct precipitation and runoff from the catchment and subtraction of lake evaporation and outflow. The model assumes no temporal lag between surface and sub-surface flow to the lake and no distinction is therefore made between surface and sub-surface components of catchment runoff (Vassiljev et al. 1998). In the absence of measurable surface drainage, lake water outflow is defined as a percentage of the overall lake volume lost through sub-surface seepage. The model calculates lake level (depth or elevation) and surface area from the derived volume and thereby requires hypsographic delineation of lake bathymetry and basin topography.

For monthly time steps, the hydrologic model simulates direct precipitation over the lake from monthly precipitation (m month^{-1}) inputs and the surface area of the lake defined by the volumetric curve. Evaporative loss from the lake is calculated with the empirical mass transfer formula of Brutsaert (1982):

$$E = N\, u \left(e_{\text{s-w}} - e_{\text{a-a}}\right) \tag{3}$$

$$N = 3.367 \times 10^{-9} \left(A^{-0.05}\right) \tag{4}$$

where u is the wind speed (m month^{-1}) over the lake, $e_{s\text{-}w}$ is the saturation vapor pressure at the surface water temperature, $e_{a\text{-}a}$ is the actual vapor pressure at the overlying air temperature, and A is the surface area (m^2) of the lake. Saturation vapor pressure (millibars) is calculated from the equation of Murray (1967):

$$e_{\text{s-w}} \; and\,/\,or \; e_{\text{s-a}} = 6.108 \,\text{EXP}\left(\frac{17.27\,T}{T + 237.7}\right) \tag{5}$$

where T is the temperature (°C) of the lake water surface ($e_{s\text{-}w}$) or overlying air ($e_{s\text{-}a}$). The actual vapor pressure of the overlying air ($e_{a\text{-}a}$) follows from the saturation vapor pressure equation:

$$e_{\text{a-a}} = e_{\text{s-a}} \times h \tag{6}$$

where $e_{s\text{-}a}$ is the saturation vapor pressure at the air temperature and h is the relative humidity expressed as a fraction.

Potential catchment evapotranspiration (m month^{-1}) is derived empirically from the equations of Thornthwaite (1948):

$$PET = 0.016 \times f \times \frac{(10T)^a}{H} \tag{7}$$

$$H = \sum_{1}^{12} \left(\frac{T}{5}\right)^{1.5} \tag{8}$$

where f is the adjustment factor related to hours of daylight and latitude, T is the mean monthly air temperature between 0° and 26.5°C, H is the heat index, and a is a cubic function of the heat index:

$$a = 0.49 + 0.179H + 7.71\times10^{-5}\,H^2\,6.75\times10^{-7}\,H^3 \tag{9}$$

Adjustment factors related to hours of daylight and latitude appear tabulated within Thornthwaite and Mather (1957). At temperatures above 26.5°C, potential evapotranspiration follows a hyperbolic function:

$$PET = -0.41585 + 0.03224T - 0.00043T^2 \tag{10}$$

where T is the mean monthly air temperature.

The rate of actual evapotranspiration (AET) is determined by potential evapotranspiration, the total area of the catchment covered with vegetation, and the availability of water stored within catchment soils. During wet months ($P > PET$) catchment evapotranspiration occurs at the potential rate and excess precipitation (P_e) recharges soil moisture stores:

$$P_e = \begin{cases} 0 & P \leq PET \\ P - PET & P > PET \end{cases} \tag{11}$$

Soil moisture (SM) increases to the defined maximum water capacity (MWC) of the soil as determined by soil texture and depth. If the amount of water within catchment soil stores exceeds the maximum water capacity the excess is assumed to recharge the lake as surface and sub-surface inflow (ΣI):

$$\Sigma I = \begin{cases} 0 & SM + P_e \leq MWC \\ (P_e - MWC) + SM & SM + P_e > MWC \end{cases} \tag{12}$$

No excess precipitation is generated if catchment precipitation equals potential evapotranspiration ($P = PET$). Consequently, soil moisture remains unchanged and overland flow and near-surface inflow is eliminated.

Evapotranspiration is modified downward from the potential value if catchment precipitation falls short of potential water demands of catchment vegetation ($P < PET$). The difference is assumed to be withdrawn from soil moisture stores at a rate

proportional to the potential evapotranspiration. This process proceeds until the soil moisture storage is exhausted. Soil moisture loss (SM_{loss}) is thereby calculated as a function of precipitation and potential catchment evapotranspiration:

$$SM_{\mathrm{loss}} = \begin{cases} 0 & P \geq PET \\ |P - PET| & P < PET \end{cases} \tag{13}$$

Isotopic Model Equations

The isotopic mass balance model calculates the isotopic composition of evaporative waters by the formula of Craig and Gordon (1965):

$$\delta_{\mathrm{E}} = \frac{\alpha^* \delta_{\mathrm{L}} - h_{\mathrm{n}}\, \delta_{\mathrm{A}} - \varepsilon_{\mathrm{tot}}}{1 - h_{\mathrm{n}} + 0.001\, \varepsilon_{\mathrm{kin}}} \tag{14}$$

where α^* is the inverse of the equilibrium isotopic fractionation factor ($1/\alpha$), h_n is the relative humidity normalized to the temperature of the lake surface water, δ_A is the isotopic composition of atmospheric moisture, $\varepsilon_{\mathrm{kin}}$ is the kinetic isotopic fractionation, and $\varepsilon_{\mathrm{tot}}$ is the total (per mil) isotopic fractionation ($\varepsilon_{\mathrm{eql}} + \varepsilon_{\mathrm{kin}}$).

Normalized relative humidity is calculated from the saturation vapor pressure of the overlying air ($e_{\text{s-a}}$) and the saturation vapor pressure ($e_{\text{s-w}}$) at the surface water temperature:

$$h_{\mathrm{n}} = h \times \frac{e_{\text{s-a}}}{e_{\text{s-w}}} \tag{15}$$

Atmospheric moisture (δ_A) is assumed to be at isotopic equilibrium with precipitation (Zimmerman et al. 1967; Zuber 1983; Horita 1990; Gibson et al. 1993; 1999):

$$\delta_{\mathrm{A}} = \delta_{\mathrm{P}} - \varepsilon_{\mathrm{eql}} \tag{16}$$

The isotopic mass balance model calculates the equilibrium oxygen isotopic fractionation factor (α) and the inverse of the equilibrium oxygen isotopic fractionation factor (α*) from the equation of Majoube (1971):

$$\ln\alpha = 1137\,T^{-2} - 0.4156\,T^{-1} - 0.0020667 \qquad (17)$$

$$\alpha^* = 1/\alpha \qquad \alpha^* < 1 \qquad (18)$$

where T is the temperature (degrees Kelvin) of the lake surface water. The per mil equilibrium isotopic fractionation ($\varepsilon_{\mathrm{eql}}$) of oxygen follows accordingly:

$$\varepsilon_{\mathrm{eql}} = 1000(1-\alpha^*) \qquad (19)$$

Additional non-equilibrium fractionation is introduced to the model during evaporation from the lake surface. This kinetic fractionation ($\varepsilon_{\mathrm{kin}}$) is controlled by molecular diffusion and moisture deficit ($1-h$) over the lake surface (Merlivat and Jouzel 1979):

$$\varepsilon_{\mathrm{kin}} = C \times (1-h_{\mathrm{n}}) \qquad (20)$$

where C is the experimentally-derived isotopic fractionation value of 14.2‰ for oxygen (Vogt 1976; Araguás-Araguás et al. 2000) and h_{n} is the humidity normalized to the temperature of the lake surface water (Equation 15). The total isotopic fractionation ($\varepsilon_{\mathrm{tot}}$) is calculated as the sum of the equilibrium and kinetic isotopic fractionations:

$$\varepsilon_{\mathrm{tot}} = \varepsilon_{\mathrm{eql}} + \varepsilon_{\mathrm{kin}} \qquad (21)$$

The model calculates the isotopic composition of inorganic carbonate (calcite) as a function of the isotopic composition of lake water and lake water temperature:

$$\delta_{\mathrm{C}} = \delta_{\mathrm{L}} + 1000\ln\alpha_{\mathrm{cal}} \qquad (22)$$

where δ_C is the isotopic composition of precipitated calcite, δ_L is the isotopic composition of the lake water, and 1000 ln α_{cal} is the per mil calcite-water fractionation defined by the standard equation of O'Neil et al. (1969):

$$1000 \ln \alpha = 2.78 \times 10^6\, T^{-2} - 2.89 \tag{23}$$

The equation of O'Neil et al. (1969) requires the average temperature of the lake water T (degrees K). Conversion of the calcite-water fractionation value between the PDB and SMOW scales utilizes the expression:

$$\delta_{\mathrm{SMOW}} = 1.03086\, \delta_{\mathrm{PDB}} + 30.86 \tag{24}$$

The equation relating the isotopic composition of lake water to that of biogenic carbonate (ostracode shell calcite) is defined by Xia et al. (1997):

$$\delta_{\mathrm{B}} = \delta_{\mathrm{L}} - 0.179\, T + 3.943 \tag{25}$$

where δ_B is the isotopic composition of the biogenic carbonate, δ_L is the isotopic composition of the lake water (‰, SMOW), and T is the temperature of the lake water.

Model Input Parameters

Initial model input parameters (Table 4-2) were derived from two decades of lake observation and historical climate data collected and compiled by both INSIVUMEH and the Global Historical Climatology Network (Vose et al. 1992). Air temperature and precipitation inputs were derived from the average of monthly values recorded at four sites (El Paso, Flores, Porvenir, and San Ignacio) within ~80 km of the Lake Salpetén watershed (Figure 4-2a). Average monthly windspeed and relative humidity were calculated from daily values recorded at Flores, approximately 30 km from the lake (Figure 4-2b). Discontinuous lake water temperature measurements collected between

1980 and 1999 were used to calculate a proportional relationship between mean-monthly air temperature and surface water temperature (Figure 4-3).

Lake basin characteristics (lake surface area and catchment surface area) were determined from topographic maps. Lake depth and volume utilized the bathymetric estimates of Brenner et al. (2002). Catchment forest cover and soil water characteristics (maximum water capacity) were obtained from geographic information system datasets (Zobler 1986; Bouman et al. 1993; Sader et al. 1994). Input values for the oxygen isotope mass balance model were derived from measured values of lake water (δ_L) and the combined average of measured local precipitation (δ_P) and surface and groundwater values (δ_I) and interpolated regional precipitation values (IAEA 2001).

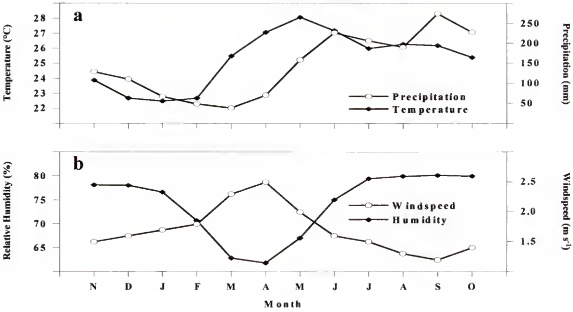

Figure 4-2. Modern climate data from the Department of Petén and western Belize. (a) Temperature (filled diamonds) and precipitation (open circles) values derived from monthly averages recorded at El Paso, Flores, Porvenir, and San Ignacio. (b) Average monthly windspeed (open circles) and relative humidity (filled diamonds) calculated from daily values recorded at Flores.

Table 4-2. Initial model input parameters for Lake Salpetén, Petén, Guatemala.

Symbol	Description	Value	Data Source
V	Initial Lake Volume	2.052×10^7 m^3	Bathymetric Map Estimate
SM	Soil Moisture Stores	0	Assumed (End of Dry Season)
P	Precipitation [a]	1755 mm year^{-1}	INSIVUMEH; Vose et al. 1992
O	Outflow (Percent of Total Volume)	1.25% month^{-1}	Deevey et al. 1980
δ_L	Isotopic Composition of Lake	4.1‰	Measured
δ_P	Isotopic Composition of Precipitation [a]	-4.0‰	IAEA 2001
δ_I	Isotopic Composition of Inflow [a]	-4.0‰	IAEA 2001
T_{Air}	Air Temperature [b]	25.3°C	INSIVUMEH; Vose et al. 1992
T_{Lake}	Lake Temperature [b]	28.5°C	Measured
h	Relative Humidity [b]	75%	INSIVUMEH; Vose et al. 1992
u	Wind Speed [b]	1.7 m s^{-1}	INSIVUMEH; Vose et al. 1992
MWC	Maximum Water Capacity [c]	0.03 m	Zobler 1986; Bouman et al. 1993
–	Catchment Surface Area	6.360×10^6 m^2	Topographic Map Estimate
–	Percent of Catchment With Vegetation	90%	Sader et al. 1994

[a] Weighted mean annual value. Inflow is assumed to equal precipitation.
[b] Mean annual value of monthly parameter varied within model experiments.
[c] Assumes an average catchment soil profile depth of 0.6 m.

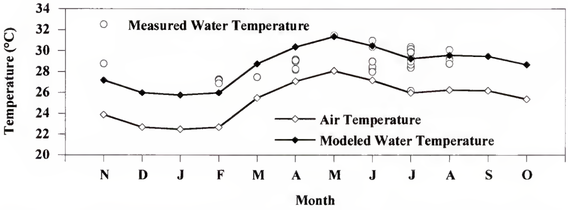

Figure 4-3. Water temperature estimates used in model simulations. Water temperature of Lake Salpetén (filled diamonds) was proportionally resolved from mean monthly air temperature (open diamonds). Discontinuous lake water temperature measurements collected between 1980 and 1999 (open circles) are plotted for comparison.

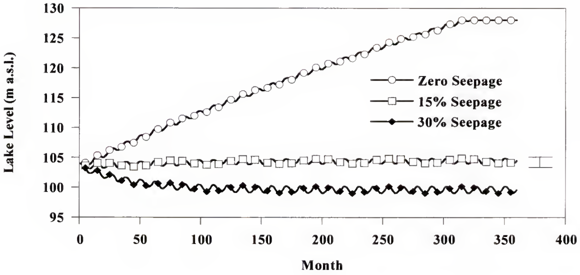

Figure 4-4. Modeled downward seepage estimates. Simulated changes in Lake Salpetén surface elevation (m above sea level) with differing rates of water loss through downward seepage. Black bars indicate the modern lake surface datum of 104 m a.s.l. (± 2 m) is shown for comparison.

Model Experiments and Results

The modern climate and catchment parameter datasets were used to approximate the seasonal and long-term variability of Lake Salpetén water levels and lake water $\delta^{18}O$. In the first set of experiments, sub-surface outflow (downward seepage) was assigned a zero value and the lake was assumed to lose water only by evaporation. This "closed basin" simulation failed to approximate modern lake levels and exceeded the spill-over height of the catchment determined from topographic maps (128 m a.s.l.). Sub-surface lake outflow was thereby progressively increased in a series of 360 month (30 year) simulations until the modern lake level was accurately predicted (Figure 4-4).

Modern lake level was achieved only with the inclusion of monthly sub-surface water losses nearing 1.3% of the total lake volume (~15% of the annual water budget). Deevey et al. (1980) identified comparable non-evaporative losses at Lakes Yaxhá and Sacnab during one year of monitoring in the early 1970s. The documented water losses,

120 cm at Lake Yaxhá and 135 cm at Lake Sacnab, represented between 15 and 19% of the total volume of the lakes. In the absence of surface outflow, Deevey et al. (1980) suggested that this lost volume was "incipient groundwater" stored for some time within clay sediments of the lake floor.

In the downward seepage configuration, the simulated lake reached hydrologic and isotopic steady-state conditions within ~15 model years or 180 monthly time-steps. The simulated lake level curve reproduced the observed pattern of seasonal lake level change (~0.9 m) with maximum stage between November and January and minimum stage between May and July (Figure 4-5a). The model experiment also reproduced the mean summer month lake water isotopic composition (+4.1‰ ± 0.3‰) and seasonal variations of ± 0.4‰ (Figure 4-5b).

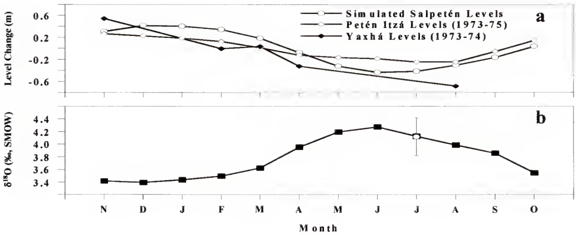

Figure 4-5. Modeled lake level and lake water $\delta^{18}O$ changes. (a) One year simulation of water level changes at Lake Salpetén (open circles) using historic data for inputs of rainfall, temperature, humidity, and windspeed. Mean monthly lake level changes for other basins in Petén (Deevey et al. 1980) are plotted for comparison. (b) Simulation of lake water $\delta^{18}O$ changes at Lake Salpetén. The mean of lake water $\delta^{18}O$ values measured during the months of May, June, July, and August is shown for comparison (open circle).

Simulated evaporation using the Brutsaert formula (Equation 3) averaged 110 mm month^{-1} and totaled ~1340 mm each year. Although no direct evaporation measurement data exist for Lake Salpetén, modeled monthly evaporation was consistent with rates of 105 and 115 mm month^{-1} reported at Libertad and San Andres, respectively (Table 4-1). Model-derived lake evaporation, however, slightly overestimated the evaporation rate of 90 mm month^{-1} measured at Lakes Yaxhá and Sacnab (Deevey et al. 1980). Modeled steady-state values of the oxygen isotopic composition of evaporated water varied seasonally between −0.7‰ and −17.5‰ (Table 4-3).

Table 4-3. Steady-state calculation of the isotopic composition of evaporated water.

M [a]	T_{Air}	T_{Lake}	h	$e_{s\text{-}a}$	$e_{a\text{-}a}$	$e_{s\text{-}w}$	h_n	α^* [b]	ε_{eql}	ε_{kin} [c]	δ_L	δ_P	δ_A	δ_E
Eq. [d]	–	–	–	5	6	5	15	18	19	20	–	–	16	14
N	23.9	27.2	0.782	29.66	23.19	35.97	0.645	0.9909	9.12	5.04	3.42	-1.90	-11.0	-10.2
D	22.7	26.0	0.781	27.59	21.55	33.52	0.643	0.9908	9.21	5.07	3.40	-1.16	-10.5	-11.7
J	22.5	25.8	0.767	27.26	20.91	33.12	0.631	0.9908	9.23	5.24	3.44	-1.21	-10.4	-11.9
F	22.7	26.0	0.707	27.59	19.51	33.52	0.582	0.9908	9.21	5.94	3.50	-0.60	-9.81	-14.1
M	25.5	28.8	0.629	32.63	20.53	39.48	0.520	0.9910	8.99	6.82	3.63	1.56	-7.43	-17.1
A	27.1	30.4	0.619	35.86	22.20	43.29	0.513	0.9911	8.87	6.92	3.96	-2.84	-11.7	-11.9
M	28.1	31.4	0.671	38.02	25.51	45.83	0.557	0.9912	8.79	6.30	4.20	-3.24	-12.0	-9.4
J	27.2	30.5	0.751	36.07	27.09	43.54	0.622	0.9911	8 86	5.36	4.28	-5.17	-14.0	-3.3
J	26.0	29.3	0.795	33.61	26.72	40.64	0.658	0.9911	8.95	4.86	4.13	-3.46	-12.4	-4.5
A	26.3	29.6	0.800	34.22	27.37	41.35	0.662	0.9911	8.93	4.80	3.99	-3.29	-12.2	-4.9
S	26.2	29.5	0.802	34.01	27.28	41.11	0.664	0.9911	8.94	4.78	3.86	-5.88	-14 8	-0.2
O	25.4	28.7	0.800	32.44	25.95	39.25	0.661	0.9910	9.00	4.81	3.55	-2.88	-11.9	-7.1

[a] The letter M denotes the month of the year.

[b] $\alpha^* = 1/\alpha$ and $\ln\alpha = 24844\,T^{-2} - 76.248\,T^{-1} + 52.612$.

[c] $\varepsilon_{tot} = \varepsilon_{eql} + \varepsilon_{kin}$.

[d] Numbers correspond to the equation listed within the model description.

Model Sensitivity

In the second set of experiments, the modern climate and catchment dataset was progressively modified to approximate the impact of increased precipitation and changes in catchment vegetation cover on the hydrologic balance and lake water isotopic composition of Lake Salpetén. Simulations of approximately two-thousand model years were completed during each experiment. Between the 400th and 800th model year, precipitation was increased or catchment vegetation cover was decreased. Precipitation and catchment vegetation levels were returned to the initial input values between the 1200th and 1600th model years. The length of each simulation was sufficient for the lake to establish steady-state equilibrium at all of the dataset parameter levels.

Precipitation over the lake and surrounding catchment was first increased to values between 105 and 130% of the modern monthly dataset. Each successive 5% increase resulted in elevation of the lake surface by nearly 1.2 m (Figure 4-6a). Incremental catchment vegetation cover decreases of 15% produced lake level changes comparable in magnitude (1.3 m) to the 5% increases in precipitation (Figure 4-6b). The principal effect of catchment vegetation loss is reduced evapotranspiration and thereby increased surface and near-surface runoff to the lake (Eqs. 11 through 13). This effect is strongest in the later wet season months of August, September, and October, when soils are at maximum water capacity and any excess precipitation is converted directly to runoff.

Step-wise precipitation increases or vegetation losses and consequent average lake level changes of ~1.25 m produced a consistent 0.15‰ decrease in the oxygen isotopic composition of lake water and followed the regression equation:

$$\delta_{\mathrm{L}} = -0.1297\left(Z_{\mathrm{ave}}\right) + 17.27 \qquad R^2 = 0.995 \tag{26}$$

where δ_L is the $\delta^{18}O$ value of the lake water and Z_{ave} is the average surface elevation of the lake (Figures 4-6c and 4-6d). Combined changes in precipitation and catchment vegetation cover increased the overall magnitude of simulated lake level and lake water $\delta^{18}O$ changes (Figures 4-7a and 4-7b).

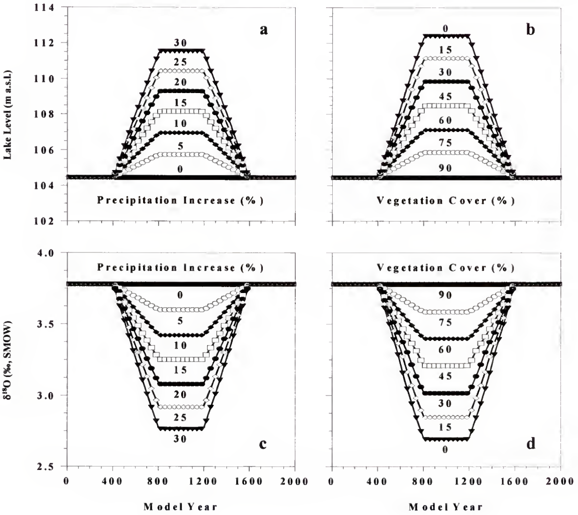

Figure 4.6. Lake level and $\delta^{18}O$ response to changes in precipitation and vegetation cover. Simulated response of lake level and lake water $\delta^{18}O$ to systematic changes in precipitation (a and b) and catchment vegetation cover (c and d). Heavy black lines delineate equilibrium lake level and $\delta^{18}O$ simulated using modern precipitation and modern catchment vegetation cover (90%).

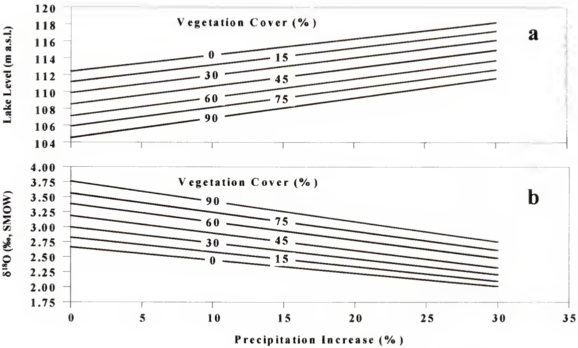

Figure 4-7. Combined precipitation and vegetation cover change effects on lake level. Simulated lake level increase (a) and lake water $\delta^{18}O$ decrease (b) as a function of combined precipitation and catchment vegetation cover changes.

Uncertainty of Model Variables

Significant error may be introduced to the model through poorly constrained variables such as relative humidity, windspeed, and lake temperature. The sensitivity of individual simulations to these variables was therefore determined by additional experiments. Relative humidity values 10% above modern and windspeed values 10% below modern produced lake surface elevation changes (through lowered evaporation) of ~0.9 m and lake water $\delta^{18}O$ decreases of 1.1 and 0.3‰, respectively (Table 4-4). The effect of relative humidity on lake water isotopic composition (δ_L) was amplified largely through the calculation of the $\delta^{18}O$ of evaporating water (δ_E) (Equation 14).

Lake water temperatures 5% below modern resulted in elevation of the lake surface (again through lowered evaporation) by ~0.3 m and reduction of lake water $\delta^{18}O$ by

~0.4‰ (Table 4-4). However, temperature fluctuations are assumed to be minor during the Holocene and variations in the isotopic composition of lake water are thought to have been dominated by changes in the relative rates and isotopic composition of hydrologic inputs and outputs.

Calculated lake water $\delta^{18}O$ may also be sensitive to changes in the isotopic composition of precipitation (Table 4-4). No long term averaged data exist for the isotopic composition of rain falling directly on the lake and some error may therefore be embedded within the interpolated δ_P values. Trace element (Mg and Sr) concentrations in sediment cores from Lake Salpetén parallel $\delta^{18}O$ records, suggesting that $\delta^{18}O$ values reflect watershed hydrology rather than changes in the isotopic composition of rainfall (Rosenmeier et al. 2002b).

Table 4-4. Sensitivity analysis of poorly constrained climate variables.

Parameter	Change	Model Result [a, b, c]
h	1% increase	δ_L decreases by 3.0%; lake level increases by 0.1 %
h	10% increase	δ_L decreases by 30.0%; lake level increases by 1.0 %
u	1% decrease	δ_L decreases by 0.8%; lake level increases by 0.1 %
u	10% decrease	δ_L decreases by 9.0%; lake level increases by 0.8 %
T_{Water}	1% decrease	δ_L decreases by 2.0%; lake level increases by 0.3 %
T_{Water}	5% decrease	δ_L decreases by 10.0%; lake level increases by 0.3 %
δ_P	1% decrease	δ_L decreases by 1.0%; lake level remains unchanged
δ_P	10% decrease	δ_L decreases by 8.0%; lake level remains unchanged

[a] Annual average at steady-state (equilibrium condition) is listed.
[b] Any 10.0% decrease in δ_L is equivalent to ~0.35‰ (SMOW).
[c] Any 1.0% increase in lake level is equivalent to ~1.0 m.

Discussion

Declining biogenic carbonate $\delta^{18}O$ values are preserved in sediment core records from Lake Salpetén between 3300 and 2400 cal yrs B.P. (Figure 4-8a) coinciding with palynologically documented forest loss (Figure 4-8b) (Rosenmeier et al. 2002a). Minimum $\delta^{18}O$ values, suggesting high water levels occurred between 2400 and 1800 cal yrs B.P. Independent evidence for high water level is provided by radiocarbon dates on aquatic gastropod shells found in exposed lake sediments that lie between 1.0 and 7.5 m above present lake surface (Rosenmeier et al. 2002b). The oldest dates indicate that the lake filled to a depth of more than 7.0 m above modern lake stage by ~2800 cal yrs B.P., but most indicate episodes of high lake level between 2000 and 1200 cal yrs B.P. These ages coincide with reduced $\delta^{18}O$ values and support the inference for increased water input to Lake Salpetén. Forest removal may have decreased evapotranspiration and soil moisture storage in the watershed, thereby increasing catchment water yield and transport of isotopically light surface and ground waters to the lake. Increased delivery of meteoric waters would increase lake volume, decrease the proportion of the hydrologic budget lost to evaporation, and decrease lake water $\delta^{18}O$. Alternatively, high water levels and $\delta^{18}O$ minima may reflect increased precipitation and reduced evaporation beginning after 3300 years ago.

Simulated late Holocene lake level elevations of ~7.0 m and consequent lake water $\delta^{18}O$ changes necessitate either sustained precipitation increases of 30% relative to modern values or reduction of vegetation cover to less than 15% of the total watershed area. Any substantial precipitation increase, however, is incompatible with late Holocene aridity inferred from undisturbed sediment core sites in the northern Yucatán Peninsula

(Hodell et al. 1995; 2001) and the Caribbean (Hodell et al. 1991; Haug et al. 2001; 2003). These records indicate a trend toward drier climate conditions beginning ~3300 cal yrs B.P. and intensifying after 2800 cal yrs B.P. Most sediment core records from circum-Caribbean sites instead indicate that the wettest period throughout the Northern Hemisphere tropics occurred between 7000 and 5000 years ago. Hodell et al. (2000) suggested that this period of increased precipitation was related to an increase in the intensity of the annual cycle driven by orbital parameters.

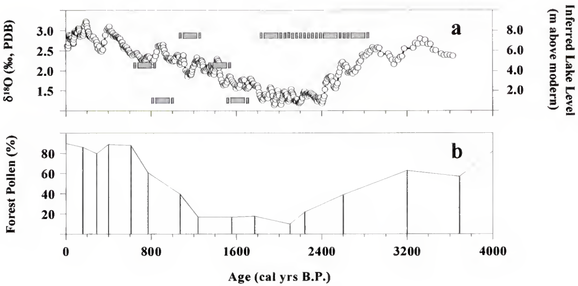

Figure 4-8. Lake Salpetén sediment core records. (a) Oxygen isotopic composition of ostracode valves (open circles) in Lake Salpetén sediment cores and lake level high stands inferred from subaerial lacustrine deposits ~1.0 to 7.5 m above modern stage (Rosenmeier et al. 2002a) versus calendar age. Oxygen isotopic data were smoothed with a 5-point running mean to illustrate long-term trends. Age of the lake level high stands was determined by AMS ^{14}C dating of aquatic gastropods (Rosenmeier et al. 2002b) and corrected hard-water lake error. (b) Relative abundance of high forest pollen taxa from Lake Salpetén sediment cores (Leyden 1987; Islebe, unpublished data). High forest taxa include species of the families Moraceae and Urticaceae.

Models of the coupled ocean and atmosphere response to middle Holocene orbital forcing show enhancement of precipitation in parts of Central America, including the Yucatán, largely as a result of northward migration of the ITCZ (Harrison et al. 2003). These simulations suggest a precipitation increase of only about 10% annually and fall short of the 30% increase dictated by Lake Salpetén model experiments. Thus, low lake levels and high $\delta^{18}O$ values inferred for the early Holocene at Salpetén were likely controlled, in part, by vegetation coverage.

Later Holocene high lake level and minimum lake water $\delta^{18}O$ as a result of substantially reduced catchment vegetation cover is consistent with palynologically documented forest loss at Lake Salpetén (Figures 4-9a and 4-9b). Model simulations incorporating pollen-derived deforestation rates reproduce much of the trend in the sediment core biogenic carbonate $\delta^{18}O$ record and explain nearly 75% of the observed variance (Figures 4-9c and 4-9d). Nonetheless, these simulations do not encompass the full magnitude of measured sediment core $\delta^{18}O$ values and fail to detect abrupt increases centered at 1500, 1100, 900, 500, and 200 cal yrs B.P.

Minimum oxygen isotopic values coincide with deposition of the *Maya clay*, erosional sediments attributed to human-induced deforestation of the Salpetén catchment (Brenner 1994). Soil erosion and decreased soil moisture storage may have accelerated catchment water yield and the inflow of isotopically depleted waters to the lake. Additional experiments assuming reduced soil thickness (i.e., soil erosion) and decreased water storage capacity, however, accounted for no more than ~0.1‰ of the 0.4‰ difference between the sediment core record and model simulation of the $\delta^{18}O$ minimum. Even if changes in the soil moisture budget were consistent with the core and model

offset, decreased catchment water storage capacity would not account for the periodic reduction of inflows inferred from the abrupt $\delta^{18}O$ increases. Changes observed in the lake core $\delta^{18}O$ record therefore necessitate either additional variability in catchment vegetation cover or more subtle regional precipitation changes.

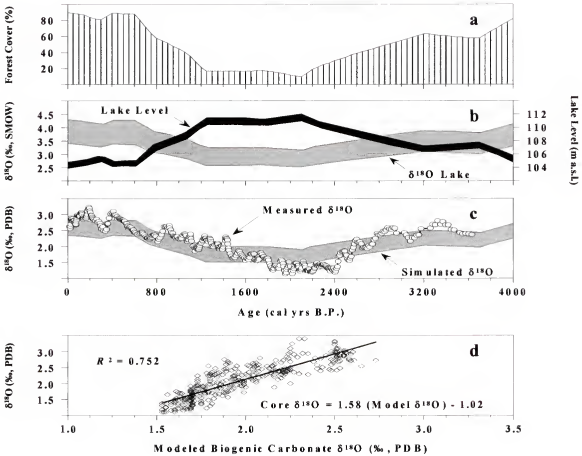

Figure 4-9. Model simulations incorporating palynologically-documented vegetation changes. (a) Pollen-based simulation of catchment vegetation cover changes and (b) consequent lake level (black line) and lake water $\delta^{18}O$ (gray line) variations. The thickness of lines results from the seasonal variations in climatic inputs. (c) Comparison of the Lake Salpetén sediment core biogenic carbonate $\delta^{18}O$ record (open circles) and simulated biogenic carbonate $\delta^{18}O$ (gray line). Model ages and sediment core ages were converted to calendar years before present. (d) Linear regression of model and sediment core $\delta^{18}O$ values.

It is conceivable that rapid vegetation changes may have altered lake hydrology while remaining palynologically-undocumented. Pollen records from Lake Salpetén are characterized by broad sampling intervals and lack the resolution to distinguish the short-term fluctuations that are discernible in the $\delta^{18}O$ profile. The full variability of the $\delta^{18}O$ record was recreated through the use of a synthetic vegetation change record characterized by a maximum decrease of forest cover to less than 5% of the total watershed area and periodic relaxations of catchment disturbance and temporary forest regrowth (Figures 4-10a and 4-10b). Some of the variability incorporated in the artificial pollen record is implausible, however, and either exceeds the known response rate of catchment vegetation and/or requires over 100% of the watershed to be forested. It is therefore likely that some of the change observed in the Salpetén $\delta^{18}O$ record reflects the additional influence of climate variations (precipitation and evaporation changes).

Minimum $\delta^{18}O$ values between 2400 and 1800 cal yrs B.P. were achieved through the superimposition of both a protracted ~15% average increase in precipitation and pollen-based estimates of vegetation cover change (Figures 4-10c and 4-10d). Abrupt changes in the $\delta^{18}O$ record were also accomplished through the superimposition of precipitation and vegetation cover changes. Simulation of the abrupt changes necessitated the rapid onset of dry conditions (precipitation decreases of at least 10% relative to modern) centered near ~3300, 2900, 500, and 200 cal yrs B.P. These model simulations also required the inclusion of particularly dry conditions between ~1500 and 800 cal yrs B.P.

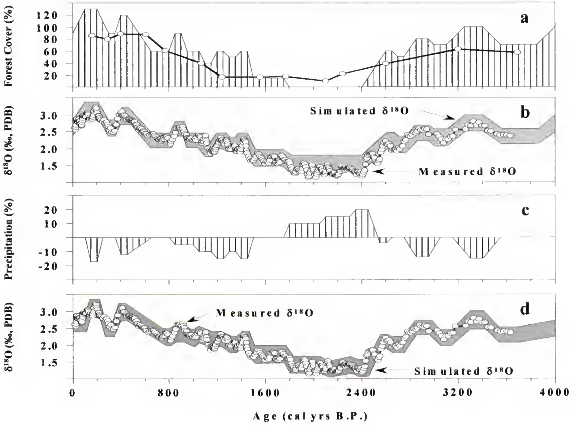

Figure 4-10. Model simulation incorporating synthetic vegetation and precipitation changes. (a) Synthetic catchment vegetation cover change record (black bars) used to simulate the variability observed in sediment core $\delta^{18}O$ values from Lake Salpetén. Pollen-based reconstruction of vegetation change (open circles) is shown for comparison. (b) Comparison of the Lake Salpetén sediment core biogenic carbonate $\delta^{18}O$ record (open circles) and biogenic carbonate $\delta^{18}O$ values (gray line) resulting from use of the synthetic vegetation record. (c) Precipitation changes (percent increase or decrease from modern) used to further alter the hydrologic and isotopic mass balance model of Lake Salpetén. (d) Comparison of the sediment core biogenic carbonate $\delta^{18}O$ record and biogenic carbonate $\delta^{18}O$ values resulting from the superimposition of precipitation changes and pollen-based simulations of catchment vegetation change.

Inferred late Holocene precipitation decreases at Lake Salpetén complement drying events recorded in northern Yucatán lake cores and suggest that shifts in moisture availability may have been regional in extent. Moreover, the precipitation increases appear coincident with Maya cultural transitions. Model-derived precipitation at Lake

Salpetén was high during the Middle and Late Preclassic periods (Figure 4-11a). These levels decreased abruptly near the boundary between the Preclassic and Classic and remained low throughout the Classic. The sediment record from Lake Punta Laguna (Curtis et al. 1996) displayed similarly abrupt $\delta^{18}O$ increases after ~1800 cal yrs B.P. and remained high throughout the Classic period, suggesting dry conditions (Figure 4-11b). In Lake Chichancanab, $\delta^{18}O$ increased abruptly at ~1200 cal yrs B.P. and marked the onset of a nearly 200 year drought in the Terminal Classic and earliest Postclassic (Figure 4-11c) (Hodell et al. 1995; 2001).

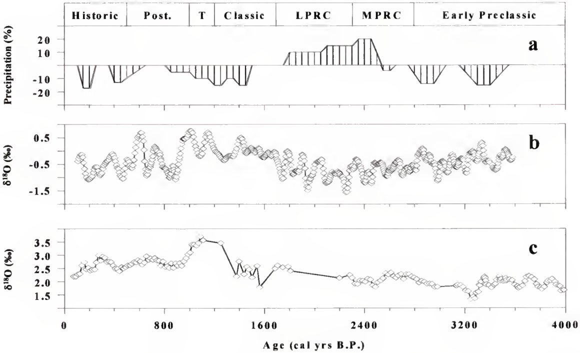

Figure 4-11. Model-inferred precipitation and lake core $\delta^{18}O$ records from northern Yucatán. (a) Regional precipitation changes inferred from the hydrologic and isotopic mass balance model of Lake Salpetén and oxygen isotope records from (b) Punta Laguna (Curtis et al. 1996) and (c) Chichancanab (Hodell et al. 1995) versus age and Maya cultural periods (Rice and Rice 1990): Post. = Postclassic, T = Terminal Classic, LPRC = Late Preclassic, MPRC = Middle Preclassic.

Conclusions

Model simulations of the hydrology and isotopic composition of Lake Salpetén indicate that nearly 75% of the measured lacustrine $\delta^{18}O$ variability may be explained by palynologically-documented changes in catchment forest cover. Although the effects of Maya deforestation complicate the interpretation of $\delta^{18}O$ data, it appears that a portion of the regional climate signal is preserved in Petén lakes. Vegetation changes alone do not reproduce the full range of variability observed in sediment core records, particularly abrupt $\delta^{18}O$ changes. Rapid vegetation changes, undetected in the relatively coarse pollen sampling interval of Lake Salpetén sediment records, may explain this offset. However, many of the required vegetation change additions exceed the potential response rate of catchment vegetation. The discrepancy between the core and model results is therefore more likely the consequence of the combined effects of changing precipitation and catchment vegetation alteration. Climatic interpretations suggest greater moisture availability between 2400 and 1800 years ago, during the expansion of Middle and Late Preclassic Maya settlement. Increased $\delta^{18}O$ values occurred as a consequence of abruptly decreased precipitation ~3300, 2900, 500, and 200 cal yrs B.P. and as a result of a period of protracted aridity between 1500 and 800 cal yrs B.P. Changes in the $\delta^{18}O$ of biogenic carbonate and inferred lake level changes in Lake Salpeten therefore resulted from human-induced deforestation and natural climate change.

CHAPTER 5
CONCLUSION

Oxygen isotope ($\delta^{18}O$) profiles from Lake Salpetén, Guatemala sediment cores indicate that human alteration of watershed hydrology confounds interpretation of $\delta^{18}O$ as a proxy of changing relation between evaporation and precipitation (E/P). Lower $\delta^{18}O$ values may represent decreased E/P (wetter climate) and/or greater surface runoff and groundwater inflow to the lake caused by human-induced deforestation. For example, declining $\delta^{18}O$ values between 3300 and 2400 cal yrs B.P. coincided with palynologically documented forest loss that may have led to increased inflow. Minimum $\delta^{18}O$ values, suggesting high water levels occurred between 2400 and 1800 cal yrs B.P.

High lake stands are also documented at this time by radiocarbon dates on aquatic gastropods found within subaerial soils (i.e., lake sediments) ~1 m to 7.5 m above the present lake stage. Following the period of minimum $\delta^{18}O$ values, the isotopic signal increases in a series of steps centered at 1500, 1100, 900, 500, and 200 cal yrs B.P. These $\delta^{18}O$ increases may reflect a series of aridity increases and/or decreased hydrologic inputs as a consequence of forest recovery associated with population declines. Most of these steps in the $\delta^{18}O$ signal correspond with discontinuities in Maya cultural evolution. Whether this coincidence represents a response of culture to climate change or a response of environment to human disturbance is not known.

Oxygen isotope records from northern Yucatán lake sediment cores provide evidence of a link between the Terminal Classic Maya collapse and declining moisture availability beginning ~1200 years ago. Lake Salpetén yields ambiguous results, most

probably because of human disturbance of watershed hydrology. Model simulations indicate that contradictory results may be due to the superimposition of climate changes and vegetation changes that altered hydrologic budgets of individual lakes and produced different $\delta^{18}O$ patterns. High-resolution paleoclimate records from other sites throughout the Maya lowlands are required to further address any potential causal relationship between drought and cultural demise.

The most appropriate isotopic records from the southern Maya lowlands will likely be found in lakes whose drainage basins were not densely settled. Archaeologists can help identify these optimal study lakes. Deciphering the paleoclimatic record from the Maya lowlands will thus involve collaboration between archaeologists and paleoenvironmental researchers. A better understanding of the long-term relationship between climate and Maya culture will not only have bearing on the interpretation of the archaeological record, but should prove informative about the future prospects for sustainable agriculture in a region that is once again becoming densely populated.

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BIOGRAPHICAL SKETCH

Michael Rosenmeier was born in the midst of a snow storm on April 17, 1972, in the bustling metropolis of Ripon, Wisconsin (population 7,111). He graduated from Ripon Senior High School in 1990 and, after briefly entertaining the idea of life as an artist, completed a Bachelor of Arts degree in anthropology from the University of Wisconsin – Milwaukee in 1994. Michael continued his academic pursuits for two years at the University of Minnesota – Duluth and studied the limnology and sedimentary history of Lake Malawi, East Africa. He transferred to the University of Florida late in 1997 where he began research on the Holocene paleoenvironmental history of Lake Salpetén, Guatemala. He will complete his Ph.D. at the University of Florida in the fall of 2003 and has accepted a tenure-stream faculty position at the University of Pittsburgh (Department of Geology and Planetary Science).

I certify that I have read this study and that in my opinion it conforms to acceptable standards of scholarly presentation and is fully adequate, in scope and quality, as a dissertation for the degree of Doctor of Philosophy.

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This dissertation was submitted to the Graduate Faculty of the Department of Geological Sciences in the College of Liberal Arts and Sciences and to the Graduate School and was accepted as partial fulfillment of the requirements for the degree of Doctor of Philosophy.

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